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EPICYCLES OF EROSION IN THE VALLEYS OF THE COLORADO PLATEAU PROVINCE

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ABSTRACT

The results of physiographic and historical studies of recent channeling in previously well-aggraded valleys in the Colorado Plateau Province are here presented. Discussion is given to the unconformities and alluvial terraces in the valley fills, evidences of other epicycles of erosion. Studies were made of naturally barren areas to determine if there had been a change in the gradational processes, which would be expected if climate had changed sufficiently to cause the accelerated erosion in adjacent valleys. Conclusions are that utilization of the region by man and the consequent reduction and modification in the plant cover are major factors in starting the new epicycle of erosion.

A new epicycle of erosion has recently begun in southern Utah and other parts of the Colorado Plateau Province. Valleys well aggraded with stream-borne alluvium are now being trenched deeply and their silts and sands swept to the Colorado River. The soil mantle which has been stabilized for geological ages on plateaus and rock platforms is now being incised in certain places by steep-sided gullies. Probably in no other region is there to be found a more widespread stream rejuvenation, originating in the last few decades.

This reversal of gradational processes in the Plateau region has been recognized by a number of travelers and investigators. Gregory and Moore,[2] who base their statements on field evidence supported

[1] Conservationist, U.S. Department of Agriculture, Forest Service.

[2] Herbert E. Gregory and Raymond C. Moore, "The Kaiparowits Region," *U.S. Geol. Surv. Prof. Paper 164*, pp. 143–44.

by detailed historical data, maintain that the present habit (degrading) of the streams in southern Utah has been established during the past 50 to 60 years. A review by Kirk Bryan[3] of accelerated erosion over large areas of the Southwest shows that the channeling is of recent origin for most of the valleys considered. An examination by the writer during 1931 to 1934 of some 115 drainages, including all of the tributaries of the Colorado and Green rivers above Lees Ferry, together with the Virgin River, Kanab Creek, and Johnson Wash, showed, with but few exceptions, that erosion in them has recently been accelerated and that most of the streams now are either scouring bedrock or rapidly removing the valley alluvium.

This accelerated erosion already has resulted in the destruction of farm and grazing lands and the depopulation of settlements. Its outstanding significance is stressed by the construction of the Boulder Dam, as the life of that huge reservoir, as well as the perpetuity of agriculture, cities, and industry dependent on the water stored, will be determined by its rate of silting. Knowledge of what caused this speeding-up of erosion is important in planning control measures and also is of interest from a scientific viewpoint.

The cause of this new epicycle, however, is not easily distinguishable. The depletion and modification of the plant cover have been ascribed as a major cause by ecologists, range investigators, and others. This is doubtless based on their experience with vegetation and knowledge of its effect upon the rate of surficial runoff and soil erosion. Engineers, accustomed to building large structures to control floods, have been more interested in endeavoring to handle accumulated water and erosion débris than in the causes. There has been a tendency of some geologists to consider the present erosion inconsequential and the result of either climatic change or diastrophism, the dominant causes of stream rejuvenation in the geological past. Their opinions doubtlessly have been predicated on the knowledge that the valleys, canyons, and plateaus, as well as their soils and sediments, are the products of erosion operating over long periods of time.

[3] Kirk Bryan, "Date of Channel Trenching (Arroyo Cutting) in the Arid Southwest," *Science*, N.S., Vol. LXII (1925).

Important contributions in the analysis of the problem and in the determination of the cause can be made by geologists. These have to do with the origin and manner of deposition of the valley alluvium; stream activity during the period of valley filling; and whether there was continuous deposition or intermittent filling and channeling up to the present epicycle. Briefly, much of the past and probably the immediate future history of a stream may be gathered from an interpretation of the profile and structure of the sediments. A comparison of sedimentation and erosion on naturally barren watersheds with that on watersheds which are more or less covered with soil and vegetation should aid in determining the causes of the present epicycle of erosion.

DEVELOPMENT OF THE TOPOGRAPHY

The manner and stage of topographical development play a large part in the erosion problem. The plateau area of southern Utah, which is typical of a much more extensive region, has been greatly influenced by the structure and composition of the rocks. The great sedimentary systems underlying the area, made up of alternating series of soft and hard rocks, predominantly sandstones of the Mesozoic system, dip gently to the north, which is opposite to the direction of flow of most of the streams. In broad outline the topography descends from the high plateaus to the Colorado River in a series of slightly tilted platforms, each terminated by a great cliff or scarp. The streams have cut their relatively narrow valleys transversely through these successive steplike features. The valleys are widened where the stream courses have excavated into the weaker Lower Cretaceous, Morrison, Chinle, and Moenkopi formations. Many of these widened sections are separated by canyons through which the streams flow with increased gradient. The valleys are widest and have their greatest depth of alluvial fill where the streams emerge from the narrow canyons upon the platforms. Here, also, the most water is available for agricultural purposes, and vegetation normally is most abundant.

The topography as a whole is in a youthful stage. However, the more active cutting which carved the valleys and canyons in solid rocks was retarded, deposition began, and the building-up of the

valley floors continued, with interruption in some places, to the present epicycle of erosion. The aggrading of the valleys is probably only a temporary stage in the general process of degradation of the province, and considered in the light of the erosion cycle is of little significance. But it is of real importance to the population dependent on the use of the land and the water, although the period of use may be brief, as gauged in terms of geologic time.

FIG. 1.—The wash in Johnson Valley, Utah, near the mouth of Skutumpah Valley. A channel 40 feet in depth, exposing a cross section of the bedded alluvium, has been cut since settlement in the previously aggrading valley floor. Recent erosion, similar in character, is occurring throughout the Colorado Plateau Province.

THE ORIGIN OF THE VALLEY FILL

The recent deep cuts often traverse the entire width of a valley, and expose the materials of the valley fill in cross section (Fig. 1). The alluvium is predominantly sand, but in places contains considerable clay and a little gravel. The sands are remarkably well stratified, laminated, and cross bedded, and small-scale ripple marks are common. Tan, maroon, red, blue-gray, and white layers, replicas of the formations from which the materials have been borne by the streams, alternate with each other in the sediments. These features are characteristic of the work of a braided stream spreading and

anastomosing over a valley floor, depositing its load in thin layers. Humic horizons in the fill mark old surfaces that were left undisturbed over a period sufficient to allow plant material to enter into the make-up of the soil. In some places peat layers, varying from a mere trace to 2 feet in thickness, occur interstratified with the sands and clays, recording the existence of meadows, swamps, and lakes at different stages in the upbuilding of the valley floor.

FIG. 2.—Old and new erosion in Sheep Creek, a tributary of Paria River. Alluvial terraces now occupied by trees several decades in age, above the recently deepened stream channel indicate prehistoric periods of accelerated erosion.

The manner in which many sections of the valleys were aggraded can be observed in the few remaining unchanneled valleys and flats. Even today lakes and swamps, formed by sand dunes or alluvial fans of tributaries damming the main channels and impounding the streams, occur along these intact drainages just as they did along the now channeled valleys during previous stages.

PAST EPICYCLES

Alluvial terraces along the sides of the valleys and erosional unconformities in the sediments in certain sections of many of the drainages evidence prehistoric channeling of the alluvium. These stages of channeling have been termed "epicycles of erosion" by

Gregory.[4] Terraces cut from the alluvium and left on the sides of the valleys where present indicate previous periods of degradation (Fig. 2). Such terraces are absent in most of the valleys and, if present, are only locally developed. The maximum number observed in any one valley was five, which varied in height from 2 feet to 65 feet.

The unconformities are marked by outlines of old channels that were cut into the sediments, beveling the layers of the previously

FIG. 3.—Erosional unconformities in the recently exposed fill in Kanab Creek Valley near Kanab, Utah. These unconformities, found in many places in southern Utah, show that aggradation has been interrupted by periods of cutting.

deposited materials. They now form a distinct line between the old deposits and those which later refilled the channels (Fig. 3). Mud balls surfaced with pebbles occur on some of the old refilled channel bottoms, just as they do in recently formed washes. Steep banks and well defined, undisturbed, caved material occur in the outlines of some of these old channels, indicating that the old cutting was very rapid and that refilling of the eroded channels took place soon after degradation (Fig. 4). These unconformities sometimes occur

[4] Gregory and Moore, *op. cit.*

in different horizons of a valley fill, showing that cutting and filling have been recurring processes.

DIFFERENCES IN PAST AND PRESENT EPICYCLES

The terraces and erosional unconformities occur throughout the province, although they are not found in every drainage or in all parts of a single drainage; nor are they equally developed when present. Nevertheless, they prove the occurrence of former periods of accelerated erosion. There are, however, certain differences

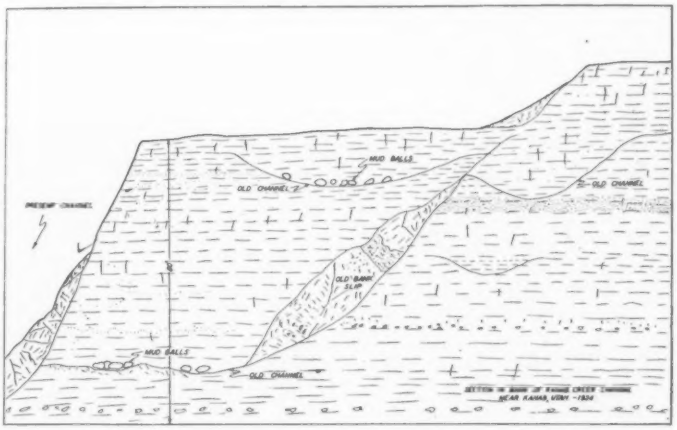

FIG. 4.—Diagram from a photograph of a recently exposed valley fill in Lower Kanab Creek channel. The unconformities, mud balls, and little-disturbed caved material indicate several stages of old channeling that were soon followed by deposition. The new cut is 80 feet deep and has occurred since settlement.

between the old and recent periods of acceleration which are of particular significance.

In the narrow valley along Kanab Creek, near the town of Kanab, Utah, for illustration, several old erosion channels are exposed in the banks of the recently formed gorge which is 90 feet deep and from 200 to 300 feet wide. There are also two terraces, one 62 feet and the other 120 feet, above the present valley floor. The period of aggradation, during which the valley floor was built up to the level of the upper terrace, was interrupted by shorter periods of

stream erosion, during which channels were cut in the sediments, and then subsequently filled. From the level of the upper terrace the stream incised itself into the valley in three stages. The greater severity of the recent cutting is indicated by the fact that no cases were found where the old channels had cut to as low a level as the present channel bottom.

In the upper reaches of the same drainage a broader valley, eroded in the lower Cretaceous formations and subsequently filled

FIG. 5.—One of the recently formed washes in Upper Kanab Valley. The absence of unconformities exposed in the alluvium, lack of terraces above the present valley floor, and the presence of layers of humic soil and peat in the sediments signifies a long period of aggradation prior to the recent cutting. The upper valleys of many drainages in southern Utah were not channeled during previous epicycles of erosion, as they are today.

with sands and clays from the Mesaverde and Wasatch formations, is now being deeply channeled by several washes working back by headward erosion (Fig. 5). Here however, there are no terraces to mark higher stages of valley filling nor erosional unconformities in the alluvium. It is evident that upper Kanab Valley is now being channeled for the first time since the beginning of the aggrading cycle.

Johnson Valley, 12 miles to the east, and comparable in most respects to Kanab Valley, has neither terraces nor significant unconformities exposed in its recently cut channel, which extends 20 miles or more upstream, except for a few traces of shallow old channels near its mouth. Kitchen Valley, a tributary of the Paria, is being cut by a steep-sided wash 40 feet deep, which exposes previously undisturbed sands and clays. The present level of its valley is probably at the maximum height attained by aggradation, there being no alluvial terrace along its margins. Thus in both of these valleys, as well as in Upper Kanab, the fills are being extensively and deeply eroded for the first time. Many similar illustrations can be cited.

The channeling during previous epicycles seems to have been confined largely to the narrower parts of the valleys, many of which are bordered by areas of bare rock, or to sections where the valleys are entered by tributaries that drain steep country. It did not extend into the upper valleys of these drainages. The broader valleys bordered by more subdued, more completely vegetated watersheds, in most cases, are now being channeled for the first time. Channels have been cut to lower levels and entire valleys that have not previously been channeled, as well as uncut sections of other valleys, are now extensively eroded.

CAUSES OF THE EPICYCLES OF EROSION

Present accelerated erosion, as well as the past epicycles, may have been caused by a change in one or more of the factors, which for long durations have maintained the aggrading or graded condition of the runoff, whether as drainage from the tributary slopes or as streams in the main valleys. The overthrow of the balance between erosion and soil accumulation or sedimentation could have resulted from diastrophism increasing the gradient of the streams, from increased quantity or intensity of precipitation, or from the destruction and modification of the plant cover, or combinations of these factors.

Diastrophism.—Increase in the gradient of valleys and slopes has been suggested as a possible cause of the recent acceleration, but cutting of equal magnitude on different drainage systems, which

traverse the region in all directions, would require such a delicate adjustment of uplift that diastrophism as a cause does not seem tenable.

Climatic changes.—The climate of the region is erratic, with extremes locally as well as regionally. There is a variation from cool and humid in the high plateaus and mountains to hot and arid in the lower interior areas. The character of the individual storms often plays a greater part in erosion than the total annual precipitation.

FIG. 6.—Bryce Canyon, Utah. The deposits at the base of this amphitheater, which is naturally barren of vegetation, do not show a change in the gradational process to indicate there has been a change in climatic condition that would account for the recent epicycle of erosion in the region.

Changes in climate influence the gradational processes in a variety of ways. A gradual decrease in precipitation might cause degrading streams to begin to aggrade their valleys. A sudden decrease of rainfall severe enough to quickly reduce the plant cover might accelerate erosion, especially if the intensity of individual storms were not diminished. A moderate increase in rainfall, on the other hand, may result in a better vegetative cover on the slopes and in the valleys and a consequent decrease in erosion. Greatly increased precipitation usually causes aggrading streams to degrade their valleys, at least for a time. Much depends, also, upon the intensity

of rainfall. A change involving an increase in rate of rainfall during individual storms would doubtless accelerate erosion, even if no increase in total annual precipitation occurred. A decrease in rainfall intensity would have the opposite effect.

In the Colorado Plateau Province climatic records are so fragmentary, so restricted in distribution, and generally cover such a short period, that they do not supply evidence of climatic change.

FIG. 7.—The alluvial fans below the Chinle badlands in Arizona, a naturally barren area, have no earmarks of recently accelerated erosion.

Among the best indicators of climatic change in a region are the changes in the gradational processes. In using this measure, however, it is first necessary to eliminate other possible factors, such as the reduction of gradient by aggradation, the accumulation of a soil mantle, and the presence of vegetation. This may be accomplished by observing what has happened in connection with barren areas—areas not under the control of vegetation—where it is safe to assume there has been no material change in gradient or in mantle during the past hundred years or more. Bryce Canyon in Utah (Fig. 6) and the badlands near Chinle, Arizona (Fig. 7), afford opportunity for such observations.

Bryce Canyon, comprising a drainage of over 4 square miles, is an amphitheater of fantastically eroded cliffs and canyons carved from

the pink ledges of the Wasatch formation on the eastern edge of the Paunsaugunt Plateau. The surface of the plateau slopes to the westward and away from the very rim of the amphitheater so that only the rain that falls in the unvegetated drainages constitutes the runoff from Bryce Canyon. At the lower end of the amphitheater the several canyons converge to form a valley where débris from the Pink Cliffs is deposited. Aggradation of the valley at this point is very pronounced because the change in gradient is abrupt. If the barren cliffs and gorges of Bryce Canyon had received an increase in rainfall in recent years equal to that necessary to cause the accelerated erosion in the neighboring valleys, the resultant greater runoff would have cut into the deposits at the lower edge of the amphitheater and would have transported débris farther into the valley. This change does not appear to have occurred. There has been no conspicuous abnormal channeling of the deposits, and this section of the valley is still aggrading the same as in the past. The runoff from Bryce Canyon, on reaching the lower gradient of the valley, divides into several channels which shift their courses over the valley, from time to time, during deposition. These recently formed channels are no deeper than old abandoned ones in the same area. The relative ages of these channels can be determined from the ponderosa pine trees growing in many of them. Judging from these observations it appears that the runoff was spreading, depositing, and channeling during the past several hundred years in very much the same manner as during recent years.

South of Chinle, Arizona, is a rather extensive badland area, eroded from the Chinle formation, which is barren of vegetation and receives no drainage from the vegetated upland. Here, as in the case of Bryce Canyon, no evidence of abnormal runoff could be found. These observations in both Bryce Canyon and the Chinle area indicate that climatic changes sufficient to disturb the gradational processes in the region have not occurred.

Destruction and modification of the vegetative cover.—Depletion and modification of plant cover and accompanying loss of plant litter, as stated previously, have been suggested as an important cause of more rapid surface runoff and acceleration of erosion. One investigation on the Wasatch Plateau—a part of the Colorado Plateau Province—shows that surface runoff was reduced 64 per cent and

soil removal 54 per cent on an experimental 11-acre watershed because of restoration of the previously depleted vegetation from 16 per cent to 40 per cent of a complete plant cover.[5] A study of floods and accelerated erosion in the Wasatch Mountains in northern Utah indicates that torrential runoff from portions of watersheds recently denuded by over-grazing and fire caused channeling of valley fills and Bonneville deltas that had been undisturbed throughout the post-Bonneville period.[6]

The depletion and modification of plant cover, together with other uses of land incident to settlement and ranching in southern Utah and other parts of the Colorado Plateau Province, have been sufficient to suggest the probability that these changes constitute a major cause of the present cycle of erosion. Investigators familiar with that region and its history recognize generally that widespread destruction and modification of plant cover by over-grazing has occurred throughout the whole area. It is highly significant that accelerated erosion has begun since settlement of the various localities. Gregory, who probably knows the Plateau better than any other geologist, gives many dates of settlement and the beginning of channeling which emphasize this relationship in his description of the Kaiparowits region of Utah.[7] He also states that

> the effect of grazing, especially overgrazing, on run-off and stream trenching is obvious. The net result of the interference by man with the delicately adjusted streams is an increase in amount of surface water and number of intermittent springs, accompanied by a destruction of forage and of arable land.[8]

Kirk Bryan also shows post-settlement origin for most of the valleys considered by listing the dates of channeling and periods of settlement over large areas of the Colorado Plateau Province.

The dates of settlement of different sections of the province and the times of channel inception show a sequence that immediately suggests cause and effect. St. George, the first settlement on the Colorado River drainage in Utah, was established in 1861. Herds,

[5] C. L. Forsling, "A Study of the Influence of Herbaceous Plant Cover on Surface Run-off and Soil Erosion in Relation to Grazing on the Wasatch Plateau in Utah," *U.S. Dept. Agric. Tech. Bull. 220* (1931).

[6] Reed W. Bailey, C. L. Forsling, and R. J. Becraft, "Floods and Accelerated Erosion in Northern Utah," *U.S. Dept. Agric. Misc. Publ. 196* (1934).

[7] Gregory and Moore, *op. cit.*, pp. 27–32.

[8] *Ibid.*, p. 144.

however, were moved onto Santa Clara Creek and north to the mountain meadow country during the next two years, because of the "necessity of finding more feed for the increasing livestock." Eight years later (1870) the mountain meadows on both the Virgin River and the Great Basin drainage began to erode vigorously. Kanab and Long valleys were settled in 1868, while channeling began in 1883. Paria, a settlement on the middle section of Paria River drainage, was settled in 1871, and has since been abandoned because of the destruction of the arable land by floods which began

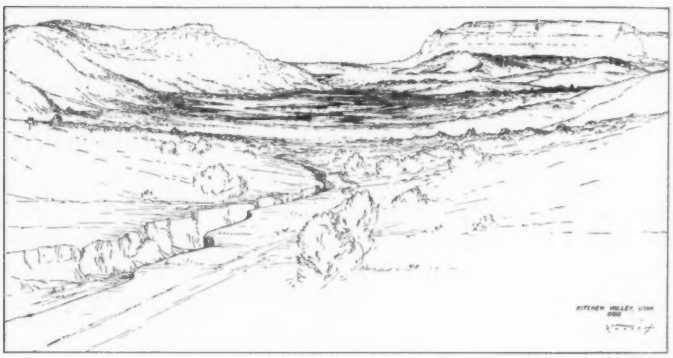

FIG. 8.—Recent erosion in Kitchen Valley, Utah

in 1883 and 1884. The settlement on the Upper Paria was established between the years of 1875 and 1878. Beginning in 1890 floods have destroyed much land and made difficult the maintenance of both roads and irrigation canals. Escalante, farther east on the Escalante River, was settled in 1875, and 15 years later channeling of the valley flats was in progress. Kitchen Valley, a tributary of the Paria, only 15 miles above the abandoned town of Paria, was converted into cattle ranches much later, but since 1915 a new channel in the valley fill, in places 40 feet deep, has cut through half the length of the valley, destroying shallow lakes and draining much meadow land (Fig. 8).

The Colorado Plateau has been used principally for grazing sheep, goats, and cattle, and for local farming. Man with his herds has penetrated to all parts of the region in search of grass and water.

Ever since original settlement the forage has been diminishing so that the plant cover has been reduced to an appalling degree. Deforestation and fire have also played some part in the process.

Investigations on widely separated areas of the open public domain in the Colorado River drainage in 1931 and 1932 showed that the plant cover had been reduced from 33 to 60 per cent on grazed areas as compared to the few ungrazed or lightly grazed areas.[9] Other Forest Service reports show the heavy reduction in plant cover by over-grazing, fire, and excessive logging.[10] Not only has the total plant cover been reduced, but the tufted, fibrous-rooted grasses have been partially replaced by single-stemmed and tap-rooted plants which are less effective in regulating runoff and controlling erosion (Fig. 9).

The valleys have been used most intensively. They were developed as settlements and ranches because of the presence of water and more abundant forage. Much of the vegetation bordering the streams was destroyed because of its ready accessibility and the necessity during early days of keeping the stock close to settlements to prevent losses by marauding Indians. Eventually the rougher and more inaccessible areas were also used for livestock production, which, in turn, was followed by depletion of the plant cover. The valleys also were common travel routes for men and livestock and became rutted with trails and wagon roads. The land near the ranches and settlements was cultivated and ditches were dug to irrigate the alluvial flats.

Before settlement many valleys of the region were being gradually built up and many of the streams were at grade. The soil-covered slopes had become relatively stabilized over large areas. However, in a topographically young country, with erratic climatic conditions, this balance between soil and plant cover and erosion is a delicate one. It is easily conceivable that with the destruction and modification of vegetation on tributary slopes and the development of gully systems in the soil, the runoff from torrential storms has been sufficiently intensified (as compared to the normal) to greatly in-

[9] Unpublished data gathered by George Stewart and Milo H. Deming of the Forest Service, U.S. Department of Agriculture.

[10] "A National Plan for American Forestry," *Doc. 12*, 73rd Cong., 1st session, pp. 433–36.

crease erosive power and thus break this balance. Wagon roads, trails, and ditches also have collected and confined runoff that ordinarily would have spread over the flat valleys. This accumulation of runoff has contributed in a major degree to the formation of channels.

How vegetation on the valley floors served to regulate runoff is illustrated by a flood observed in 1933. This flood originated on a

FIG. 9.—An over-grazed, eroding hillside and valley on Awapa Plateau in southern Utah. The vegetation was formerly chiefly a grama grass sod. The former cover has been reduced and scattered rabbitbrush has largely replaced the grass as the result of heavy over-grazing. Sheet and gully erosion are now prevalent. This condition prevails over large areas in the Southwest.

relatively barren portion of the watershed of a tributary of Upper Kanab Valley. On reaching the lower gradient at the edge of the valley the coarser débris of the stream load was deposited over an alluvial fan. The sands and clays were carried out over the nearly flat floor of the valley, where the low gradient and dense plant cover so decreased the velocity of the water and so spread it that the finer material was deposited. The rush of the flood was spent in a few minutes near the mouth of the tributary where it originated, but water flowed over the valley flats for several hours, depositing thin layers of silts and irrigating the vegetation. Lower down the valley,

however, water collected in a road and formed a stream which cascaded into a cut wash 30 feet deep that had recently been cut by previous floods. The runoff from this single flood in 1933 lengthened that channel 15 feet.

The effects of the decrease in vegetation on tributary slopes is impressively illustrated in Bryce Creek valley, a few miles below that section near Bryce Canyon discussed previously. The vegeta-

FIG. 10.—This tableland having a previously graded soil-covered slope is beginning to erode as the result of close utilization of the vegetation by the Navajo sheep and goats.

tive condition on tributary slopes and the resultant changes in gradational processes in the lower section are in marked contrast to the section near Bryce Canyon. The lower valley is bordered by Cretaceous hills with more subdued slopes, having more or less varying amounts of vegetation. Residual plant clumps and remnants of soil horizons in protected areas indicate a recent depletion of a previously more extensive plant and soil cover. This depletion is confirmed by statements of old residents, who remember a better vegetative cover in earlier days when the range had a much greater grazing capacity. As a result of the change, accelerated runoff has gullied the slopes and channeled the tributary drainages. It is

evident that this part of the valley is receiving not only the normal runoff from the barren Bryce Canyon area, but also an increased amount from the tributary slopes, which are channeling the valley and transporting the débris to the Paria River.

South of Chinle, Arizona, is another example. Here is an extensive elevated platform supported by a more resistant member of the Chinle shales. Its surface is a graded soil slope formed by erosion as the overlying Chinle and Navajo formations retreated (Fig. 10). The graded condition of this slope has been maintained long enough for

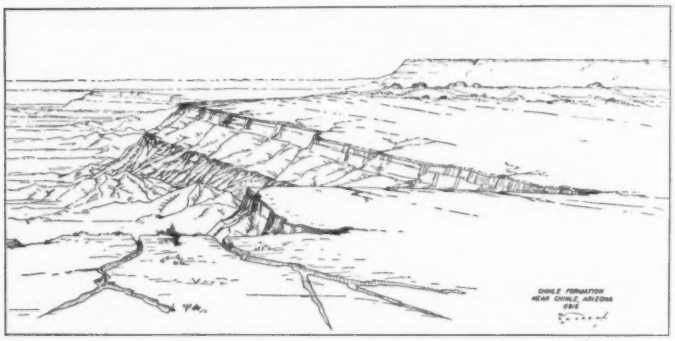

FIG. 11.—Diagram of a platform in northern Arizona, on which the soil and mantle, several feet in depth, has been undisturbed for geological ages but has recently begun seriously to erode. The vegetation on the platform has been heavily over-grazed during the past several decades.

erosion to strip back the overlying formation for a number of miles, while the Chinle drainage during this interval has carved a valley 6 miles wide in the platform. The tableland breaks off into the barren badland on three sides. Recently the graded condition of the tableland has been disturbed. Gullies are being eroded into the soil at a rapid rate, and sheet erosion and deflation have become accelerated (Fig. 11). For the first time since the tableland was formed the graded condition of the soil-covered slope is now being disturbed, apparently by accelerated runoff due to the destruction of vegetation by the sheep and goats of the Navajo Indians.

CONCLUSION

The cause of the earlier epicycles of erosion in the valleys of southern Utah and other parts of the Colorado Plateau Province

may never be determined. They may have been due to climatic change or to the destruction of vegetation. The influence of climate on gradational processes is so evident that it is always a possible cause. It is more probable that these earlier cycles were caused by vicissitude within a climatic epicycle rather than by a change in the normal climatic trend. This is indicated by the more restricted occurrence and the less well-developed cycles of the earlier erosion as compared with the present more general accelerated erosion. The apparent lack of synchronism in the cutting of the old channels, which makes impossible the correlation of erosion in different drainages, also suggests very strongly a local rather than a regional cause.

The hypothesis of destruction of vegetation is less readily accepted as a cause of the earlier epicycles, although deer or other herbivorous mammals might have become so numerous at times as to have destroyed enough of the vegetation for accelerated erosion to follow. Such conditions are believed to have existed in historical times and have even occurred in recent times, as in parts of the Kaibab Forest.

The destruction and modification of the plant cover by overgrazing appear to be the dominant causes of the recent epicycle of erosion. It is the one known change capable of reversing gradational processes that has occurred throughout the province. The formation and maintenance of soil on the slopes and of alluvium in the valleys, in a topographically young region having a variable climate, are accomplished by a delicate adjustment between the factors that control gradation. When the plant cover—one of the controlling factors—was materially reduced and modified, this adjustment was disturbed, resulting in more intensive runoff from the slopes and down the valleys, which eventuated in a reversal in the gradational process. No doubt fitful rainfall, the nature of the topography, and the character of the alluvium have contributed to the rapidity with which the erosion has progressed. The lack of any change in the gradation of naturally barren areas, such as Bryce Canyon and the Chinle badlands, evidences that there has been no effective change in quantity or rate of rainfall. The sequence of settlement, with its attendant heavy utilization of the vegetation, and the inception of channeling of the valleys, strongly support the conclusion that destruction and modification of the plant cover are major causes of the present epicycle.

THE PROBLEM OF THE MARTIC OVERTHRUST AND THE AGE OF THE GLENARM SERIES IN SOUTHEASTERN PENNSYLVANIA

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ABSTRACT

The Glenarm series of schists, marbles, and quartzites of the Atlantic Piedmont has been assigned to the pre-Cambrian in numerous recent papers. The apparent superposition of these rocks on strata of known Paleozoic age near the northwestern margin of the Piedmont province is explained by the Martic low-angle overthrust fault. It is the purpose of this paper to reopen the Glenarm-Martic problem by presenting evidence that suggests, first, that the Glenarm Series in southeastern Pennsylvania may be wholly or in part Paleozoic, and, second, that the Martic overthrust does not exist.

INTRODUCTION

The Martic overthrust fault was described in 1929 by Eleanora Bliss Knopf and Anna I. Jonas in a report on the areal geology of the McCalls Ferry–Quarryville district[1] in southeastern Pennsylvania and has been extended by these workers northward to New Jersey and at least as far south as Virginia.[2] Throughout this distance the fault is supposed to have carried the Wissahickon schist and associated quartzites, marbles, and slates of the pre-Cambrian Glenarm series[3] for a distance of approximately 20 miles northwestward across rocks of lower Paleozoic age. Since publication of the first discussion the Martic overthrust has appeared in textbooks of general[4] and structural geology,[5] in a Folio of the United States Geological

[1] "Geology of the McCalls Ferry–Quarryville District, Pennsylvania," *U. S. Geol. Surv. Bull. 799* (1929). For review and discussion of this paper see H. E. McKinstry, *Jour. Geol.*, Vol. XXXVIII (1930), pp. 472–73; Knopf and Jonas, *ibid.*, Vol. XXXIX (1931), pp. 166–67; McKinstry, *ibid.*, Vol. XXXIX (1931), pp 761–62.

[2] Jonas, "Geologic Reconnaisance in the Piedmont of Virginia," *Geol. Soc. Amer. Bull.* 38 (1927), pp. 837–46.

[3] Knopf and Jonas, "Stratigraphy of the Crystalline Schists of Pennsylvania and Maryland," *Amer. Jour. Sci.*, 5th ser., Vol. V (1923), pp. 40–62.

[4] C. Schuchert and P. O. Dunbar, *Textbook of Geology*, Part II (John Wiley and Sons, 1933), p. 278.

[5] Walter H. Bucher, *The Deformation of the Earth's Crust* (Princeton University Press, 1933), pp. 178–80.

Survey,[6] on the recently published geologic map of the United States and on state geologic maps,[7] and in a guidebook of the International Geological Congress.[8] The present writer has examined the type locality of the fault in the Martic Hills region (McCalls Ferry quadrangle) and in adjacent parts of the Pennsylvania Piedmont between the Susquehanna and the Schuylkill rivers. It is the purpose of this paper to reopen the Martic problem by presenting several lines of evidence that suggest, first, that the Wissahickon schist and associated crystalline rocks of the Glenarm series are (in the area discussed) Paleozoic rather than pre-Cambrian, and second that no major overthrust fault need be postulated to explain the relations of the rocks of the Glenarm series with known Paleozoic rocks along the supposed fault front.[9]

The Glenarm series consists, from bottom to top, of the Setters quartzite, the Cockeysville marble, the Wissahickon schist, the Peters Creek schist, the Cardiff conglomerate, and the Peach Bottom slate. The adjacent Paleozoic series consists of the Chickies quartzite, the Antietam and Harpers schists, a sequence of Cambro-Ordovician dolomites, limestones, and marbles,[10] and Martinsburg

[6] F. Bascom and G. W. Stose, "Coatesville–West Chester Folio," *U. S. Geol. Surv. Folio* 223 (1932).

[7] "Geologic Map of Maryland." Md. Geol. Surv. "Geological Map of Pennsylvania." Pa. Topog. and Geol. Surv. (1931). The legend of the Pennsylvania map contains the following note: "A belt of albite schist ¼ miles or more wide along the south side of Chester Valley from near Atglen to Conshokocken, which is included on the map with Wissahickon schist, is held by F. Bascom, B. L. Miller and E. T. Wherry to be of Ordovician age and to be north of the Martic overthrust. G. H. A."

[8] "Southern Pennsylvania and Maryland," *XVI Intl. Geol. Cong. Guidebook* 10.

[9] The structure and stratigraphy of the Piedmont district of southeastern Pennsylvania have long been controversial issues, and several of the lines of evidence summarized here have been presented at greater length in earlier publications, as indicated by numerous citations in the text. In a forthcoming article in the *Bulletin of the Geological Society of America* Professor B. L. Miller will review the more important literature and will present new stratigraphic evidence bearing on the problem. His conclusions, arrived at independently, are essentially the same as those advanced in this paper.

[10] The detailed stratigraphy of the Cambrian and Ordovician calcareous rocks of the Chester Valley lowland has been discussed by Stose and Jonas in "Ordovician Overlap in the Piedmont Province of Pennsylvania and Maryland," *Geol. Soc. Amer. Bull. 34* (1923), pp. 507–24. The general term "Cambro-Ordovician" is used here because the stratigraphy of the several formations involved does not materially affect the arguments advanced in this paper as to the age of the adjacent schists. Professor Miller will treat the stratigraphic aspect of the problem in the article mentioned above.

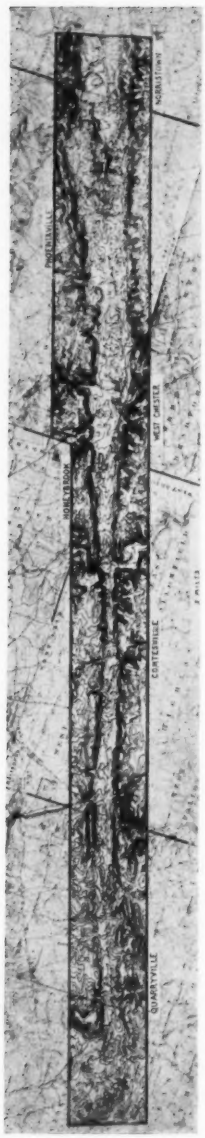

FIG. 1.—Topographic map of Chester Valley, Pennsylvania, greatly reduced from the indicated topographic sheets of the United States Geological Survey. The ridges in the Wissahickon schist at the south margin of the valley are quite as distinct as those formed by the Chickies quartzite at the north side of the valley. No longitudinal stream flows through the valley; the small southeastward flowing streams that cross it produce no embayments in the south valley side. Compare with Figures 2 and 3.

shale equivalents. Both series rest on the Baltimore gneiss. In the first part of this paper the relations of the Wissahickon schist to the known Cambro-Ordovician sequence along the supposed fault front are discussed. These relations suggest strongly that the Wissahickon overlies the Cambro-Ordovician limestone in normal sedimentary succession, and is Ordovician, probably Martinsburg, in age. In the next section it is shown that there is good ground for correlating the lower members of the Glenarm series, namely the Setters quartzite and the Cockeysville marble, exposed in anticlines south of the supposed fault front, with the Chickies quartzite and overlying limestones to the north. In a third section evidence is presented that suggests that the Peach Bottom slate, which forms the top of the Glenarm series, is Ordovician, rather than pre-Cambrian, in age.

RELATIONS ALONG THE SUPPOSED FAULT FRONT

Physiographic relations in the McCalls Ferry–Quarryville districts.—Chester Valley is a rectilinear lowland belt (Fig. 1), varying from 1 to $2\frac{1}{2}$ miles in width and extending for nearly 30 miles in a west-southwesterly direction between the Susquehanna and the Schuylkill rivers near the north edge of the Piedmont upland. The lowland is

coextensive with the outcrop of the Conestoga and other Cambro-Ordovician limestones, and its present form is evidently due to the selective lowering of these weak rocks. The north valley side is made up, throughout its length, of the Chickies quartzite, which, together with the limestone of the valley, forms the steeply dipping south limb of the Mine Ridge anticline. The Wissahickon schist forms the south valley wall (Fig. 2).

The remarkable straightness of the north side of Chester Valley is readily understandable, since the trace of the contact between two steeply dipping formations should approximate a straight line, and when, as in this case, the two formations differ markedly in resistance, should produce a rectilinear scarp. The south side of the valley is interpreted by Knopf and Jonas[11] as the eroded front of the Martic low-angle overthrust fault, the contact of Wissahickon schist and Cambro-Ordovician limestone being supposed to represent the trace of the thrust plane. A low-angle contact, whether it be a normal sedimentary contact or a fault plane, generally appears on a geologic map as in irregular line. When the two formations involved are of notably different resistance the scarp marking their contact is commonly irregular and deeply embayed, particularly where it is crossed by stream valleys. But the south side of Chester Valley is quite as straight as the north side, and this parallelism suggests that the contact between schist and limestone, whatever be its nature, dips nearly as steeply as the contact between the limestone and the quartzite.

The Chickies quartzite forms a distinct ridge, rising appreciably above the less resistant Baltimore gneiss to the north and the weak Chester Valley limestone to the south (Fig. 3). Even if the precise limits of the quartzite and limestone were unknown it is clear that this parallelism of ridge and valley would indicate a parallelism of two facies of distinctly different degrees of resistance, brought to the surface in some inclined structure. But a ridge that is identical in physiographic expression with that formed by the Chickies occurs in the schist belt south of the valley, paralleling the valley and the Chickies ridge for many miles. It appears, therefore, that some dis-

[11] Geology of the McCalls Ferry–Quarryville District, Pennsylvania," *op. cit.*, map.

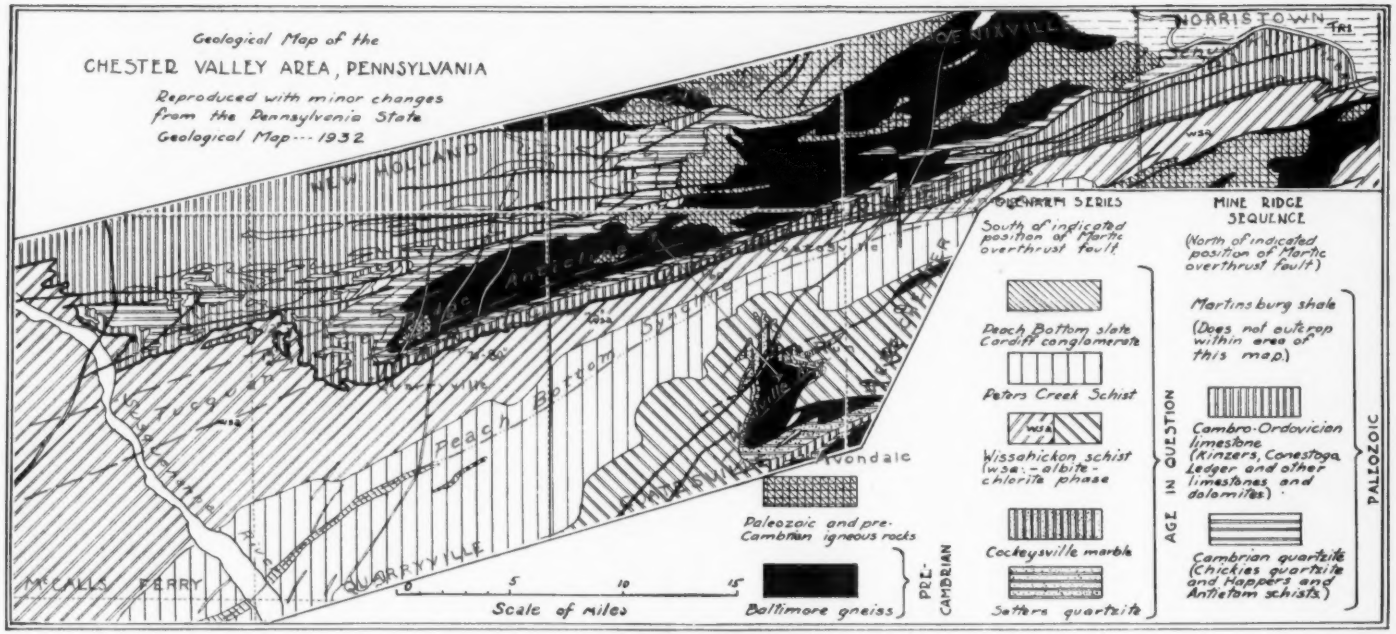

FIG. 2.—Geological map of the Chester Valley Area, Pennsylvania

tinct resistant facies within the schist is essentially parallel to the strike of the limestone and the quartzite.

The Mine Ridge anticlinal structure plunges to the southwest at a moderately low angle. The ridges upheld by the Chickies on the two limbs of the fold join to form a broad and distinctly higher area where the width of outcrop of the quartzite is increased as it swings

The Martic Hills, a highland area where the outcrop of the ridge-forming border facies of the Wissahickon is widened on the plunging nose of the Mine Ridge anticline.
The highland area formed by the widened outcrop of the Chickies quartzite on the nose of the Mine Ridge anticline.
Chester Valley, underlain by Cambro-Ordovician limestone, and limited on the north (right) by a ridge of quartzite, marking the outcrop of the Chickies quartzite, and on the south (left) by ridges of Wissahickon schist.

FIG. 3—Photograph of a block of projected profiles of the Piedmont region of southeastern Pennsylvania. View taken looking in a direction S. 70 W., along the trend of the Mine Ridge anticline. Chester Valley is etched out along the belt of Cambro-Ordovician limestone of the south limb of the anticline; the lowland north of the anticline is a part of the Great Valley. Photograph reproduced with permission of Miss Dorothy Wallace, by whom the profiles were prepared under the direction of Professor Douglas Johnson.

around the plunging anticlinal nose. This is an entirely expectable relation, widely recognized in fold mountains. The similar increase in the width of outcrop of the Cambro-Ordovician limestone at the nose of the fold, as indicated by the increased width of the subsequent lowland, is what would be expected if the limestone maintained its thickness around the nose, or, in other words, if the limestone-schist contact were essentially parallel with the quartzite-limestone contact. By far the most striking of these physiographic

relations, however, is the fact that the ridge marking the resistant border facies of the Wissahickon maintains its parallelism with the Chickies ridge to the very nose of the plunging fold, where it broadens to form an upland area, the Martic Hills, directly in line of strike with that formed by the widened outcrop of the Chickies (Fig. 3).

These physiographic relations, which certainly suggest a high-angle rather than a low-angle contact at the south margin of Chester Valley, and an anomalous parallelism in the facies of Paleozoic sedimentary rocks and the rocks of the supposed overthrust sheet, first led the writer to doubt the structural interpretation advanced in the McCalls Ferry–Quarryville Bulletin, and to make, first, a search for exposures of the critical limestone-schist contact, and secondly, a study of the structural relationships of the several formations in question.

Structural relationships in the McCalls Ferry–Quarryville district.—The limestone-schist contact is not well exposed in the McCalls Ferry–Quarryville district. In a cut of the Philadelphia and Atglen Railroad 1 mile west-southwest of New Providence (Quarryville quadrangle) the two formations are present and both are apparently involved in a minor fold, but the limestone is so thoroughly decayed that the contact relations are not clear. Neither at this point nor at the many other exposures of the schist along the south side of Chester Valley within a few tens of feet of the contact is there any appreciable brecciation, as might be expected[12] if the contact is a thrust plane of great horizontal displacement. The high dips that characterize the schist in the belt south of Chester Valley continue unchanged to the very margin of the valley, i.e., to within the zone where drag might be expected if the schist is part of an overthrust sheet. In this connection it is stated in the text of the McCalls Ferry–Quarryville Bulletin that

> the existence of this extensive overthrust fault has therefore not been established by the discordant relations seen in exposures of fault contacts. It has rather been inferred from the abnormal contacts of the pre-Cambrian Wissahickon albite schist and other pre-Cambrian formations with rocks of Cambrian and Ordovician age on the north and the northwest.[13]

[12] Knopf and Jonas point out, however, that absence of brecciation is not necessarily a valid objection to the fault theory, *ibid.*, p. 35.

[13] *Ibid.*, pp. 74–75.

The dips of the bedding of the Cambrian quartzite and of the foliation of the Cambrian schists, the Cambro-Ordovician marbles, and the Wissahickon schist correspond closely all along the southern limb of the Mine Ridge anticline, varying from 45° to 90° southward from point to point along the strike. The strike of the foliation of the Wissahickon is consistantly parallel with that of the foliation and bedding of the Cambro-Ordovician rocks. North of the village of Quarryville the strike of the Cambrian quartzite swings around the nose of the plunging fold; the westward dip, which corresponds with the structural plunge, varies from 8° to 15°. The schist 3 miles farther southwest on the nose of the fold is characterized by a marked linear mineral elongation rather than by true foliation. The strike of this linear schistosity, well shown in the railroad cut 1½ miles west of New Providence, is S. 70° W., precisely parallel with the trend of the Mine Ridge anticline in the Paleozoics. The angle of dip is essentially the same as the plunge of the Cambrian quartzites, varying from 7° to 12° to the southwest in observed exposures.

It has been pointed out above that both the schist and the Cambro-Ordovician limestone are involved in a minor fold on the nose of the anticline. Two small belts of limestone just west of the point where the main body plunges westward beneath the schist are undoubtedly due to similar minor anticlines involving both formations. The limestone belts are interpreted by Knopf and Jonas[14] as windows brought to the surface by post-thrust folding and erosion, and are held to prove that the thrust plane has a gentle dip. They do prove that the schist-limestone contact, whatever be its nature, has a gentle dip that is westward in accordance with the general structural plunge, and that it has been folded in precisely the same way as the sedimentary contacts between known Paleozoic beds directly to the east (see also page 370 of this paper).

That the Wissahickon schist has been involved with the Cambro-Ordovician series, not only in these minor folds, but also in the major Mine Ridge anticlinal structure is clearly indicated by nearly continuous exposures in the walls of the Susquehanna gorge, 10 miles west of the plunging nose discussed above. In a section[15] show-

[14] *Ibid.*, p. 75.

[15] "Lancaster County Maps," *Second Geol. Surv. Pa. Rept. of Progress CCC*, Susquehanna River Section, Sheet 3.

ing the structure of the east wall, drawn by Leslie from data gathered by Frazer, the major structure is seen to be an anticline, approximately 5 miles in breadth, called by Leslie the Tocquan (now spelled Tucquan) anticline. The Tucquan anticline is directly in line of strike with the Mine Ridge anticline, and it seems probable that the preservation of the great five-mile arch in the relatively incompetent Wissahickon schist is due to control exercised by the buried Mine Ridge core.

Alternate interpretation of the structure of the McCalls Ferry-Quarryville district.—It is evident from the marked accordance of the dip, strike, and structural plunge of the Wissahickon schist with those of the Cambro-Ordovician series that both have been affected essentially as a unit during some major period (or periods) of orogeny, the difference in behavior of the several members reflecting individual differences in competence. If it is to be supposed that the Wissahickon is a part of a thrust sheet, moved northwestward for a distance of some 20 miles, it must also be supposed that the original internal structure of the mass has been reorganized in post-thrust time so that it now parallels the structures of the rocks of the fault sole.

It appears, moreover, that the foliation of the Wissahickon schist along the south margin of Chester Valley is essentially parallel with the original sedimentary bedding of this formation. This is indicated by the parallelism of the foliation with a distinct mineralogic and lithologic banding in the schist and with intraformational boundaries in the supposed thrust sheet. It is confirmed by the parallelism of the foliation with the trend of the ridge-forming facies along the south side of Chester Valley. Since the foliation and bedding of the schist are parallel with the foliation and bedding of the several members of the Cambro-Ordovician series, it follows that if the Wissahickon is a part of a thrust sheet, it must be supposed that the bedding of the overthrust mass was essentially parallel with the thrust plane and with the bedding of the rocks of the thrust sole. Post-thrust folding might then give the observed relations. Only through such coincidences can the thrust-fault hypothesis be brought into accordance with observed facts.

The relatively simple alternate hypothesis proposed here may be

stated briefly as follows: The Wissahickon schist rests on the Cambro-Ordovician limestone in normal sedimentary succession (not necessarily in conformity). The whole sequence has been folded together during one or more periods of orogeny, the plunge of the folds being to the west at a moderately low angle. This hypothesis explains the present field relations, and from it follows as a natural and necessary consequence what is a further extraordinary coincidence if the Wissahickon is overthrust, namely the fact that the same facies of the schist (the albite-chlorite facies) is in contact with the Cambro-Ordovician limestone, and no other rock, for the whole of the distance from the Schuylkill to the Susquehanna.

Frequent use of the terms Wissahickon *schist* and Cambro-Ordovician *limestone* may have raised some doubt in the reader's mind as to the likelihood of the two rocks being the same or nearly the same age. It is beyond the province of this paper to enter into a discussion of Piedmont petrography, which has been treated at some length by Knopf, Bascom, and other competent workers. It should be pointed out, however, that the Cambro-Ordovician "limestone" of Chester Valley includes a wide variety of rock types varying from massive limestones and dolomites to coarsely crystalline impure marbles with lenses of biotite and other micas. In fact, observed variations in degree of metamorphism of the rocks in question lend strong support to the alternate structural interpretation advanced in this paper, for the writer's own observations and the published statements of Knopf and Jonas indicate that the supposed pre-Cambrian Wissahickon is (near Chester Valley) actually lower in metamorphic grade than immediately adjacent rocks known to be Paleozoic. Belts of differing metamorphic intensity, shown on a map[16] published in the McCalls Ferry–Quarryville report, cross the indicated position of the Martic overthrust as if quite unaffected by it.

Stratigraphic relations of the Wissahickon schist and Cambro-Ordovician limestone along the south margin of Chester Valley.—The hypothesis proposed in this paper implies that the schist of the south wall of Chester Valley rests on the Paleozoic limestone of the valley in normal sedimentary succession, but in the absence of good contact exposures, no direct evidence bearing on this critical point is available in the McCalls Ferry–Quarryville district. Such evidence is, however, abundantly present farther east along the line of strike of the two formations. Dr. Bascom, in the Philadelphia Folio,

[16] Knopf and Jonas, "Geology of the McCalls Ferry–Quarryville District, Pennsylvania," *op. cit.*, p. 120.

described the relations of the two formations in the Norristown quadrangle, at the eastern end of the valley as follows:

> The mica schist of the South Valley Hills appears to overlie the Shenandoah limestone without faulting or unconformity. This is indicated by the lithologic gradation between the limestone and the schist which may be seen at the northwest base of the South Valley Hills, and by the persistence of the same geodiferous quartzose beds along the contact of the two formations, and is confirmed by the fact that the outcrops show that deformation was by folding and flowage and *not by faulting*.[17]

The schist was called the Octoraro by Bascom and was assigned to the Ordovician on the basis of its normal stratigraphic position above Cambro-Ordovician limestone. In this Folio a more coarsely crystalline schist several miles south of Chester Valley was mapped as a distinct formation, the Wissahickon mica-gneiss, and was separated from the schist of the south valley side by a thrust fault. This Wissahickon mica-gneiss of the Philadelphia district was assigned to the pre-Cambrian on the basis of evidence that will be discussed below.

The contact of the crystalline limestone of Chester Valley with a calcareous facies of the schist of the south valley wall is well exposed in the cut of the Pennsylvania Railroad $1\frac{1}{2}$ miles southeast of Downingtown. In a report on the Avondale district (1916), which includes part of the Downingtown segment of Chester Valley (west of the Philadelphia-Norristown district and approximately 11 miles east of the McCalls Ferry–Quarryville district), Knopf and Jonas follow Bascom in assigning the schist of the south valley wall to the Ordovician, stating that

> The readily determined relation between the Ordovician mica schist and the great belt of Shenandoah limestone, which has been proved by fossil content to be of Cambrian and Ordovician age, establishes the age of the mica schist as Ordovician. The conformable contact of the schist upon the limestone is shown by the lithologic gradation of the limestone through a calcareous mica schist into the overlying formation, as seen in Chester Valley near Coatesville, and also by the constant presence of a single bed of geodiferous siliceous material along the line of contact, *thereby proving that the contact cannot be a fault line*.[18]

[17] "Philadelphia Folio," *U. S. Geol. Surv. Folio* 162 (1909), p. 5 (italics not in original).

[18] E. F. Bliss and Jonas, "Relation of the Wissahickon Mica Gneiss to the Shenandoah Limestone and Octoraro Schist of the Doe Run and Avondale Region, Chester County, Pa.," *U. S. Geol. Surv. Prof. Paper* 98, (1917), p. 32 (italics not in original).

In this report the higher grade schist called the Wissahickon by Bascom is again separated from the Octoraro by a thrust fault several miles south of the valley, and is assigned to the pre-Cambrian.

In the McCalls Ferry–Quarryville district Knopf and Jonas find that the albite-chlorite–schist of the south valley side, called the Octoraro in the area to the east, grades imperceptibly into a more highly metamorphic facies (the Wissahickon of the Philadelphia and Avondale districts) which occurs several miles south of the valley margin.[19] This relation evidently necessitates a change in the age assignment of either the Octoraro or the Wissahickon as previously defined. Knopf and Jonas discard the term Octoraro and include the albite-chlorite–schist in the Wissahickon formation, the redefined Wissahickon being placed in the pre-Cambrian. The Martic overthrust fault is postulated to explain the juxtaposition of the supposed pre-Cambrian schist of the south valley wall with the Paleozoic limestone of the valley.

The present writer recognizes the gradation of the albite-chlorite–schist into the more intensely metamorphosed schist to the south, and agrees with Knopf and Jonas and with Hawkins (whose work is discussed below) that the two facies actually constitute one formation. It seems, however, in view of the normal sedimentary contact of the schist of the south valley side with Paleozoic limestone in the area immediately to the east, that the enlarged Wissahickon formation should be placed in the Paleozoic rather than in the pre-Cambrian. Such age assignment would, as pointed out above, explain the striking accordance of the Wissahickon structures with those of adjacent Paleozoic rocks, and the parallelism in lithologic facies indicated by physiographic relations and field observations in the McCalls Ferry–Quarryville district.

The Coatesville–West Chester district adjoins the McCalls Ferry–Quarryville district on the east and includes the Avondale district mapped by Knopf and Jonas in 1909. In the map published in the Coatesville–West Chester Folio (1932) Bascom and Stose[20]

[19] Knopf and Jonas, "Geology of the McCalls Ferry–Quarryville District, Pennsylvania," *op. cit.*, p. 34.

[20] *Loc. cit.*

follow the usage of Knopf and Jonas in the Bulletin on the McCalls Ferry–Quarryville district in placing the albite-chlorite–schist of the south valley side in the Wissahickon and in drawing an overthrust between it and the Cambro-Ordovician limestone of the valley. This involves a reversal of the earlier assignment of the Octoraro of this district to the Ordovician[21] and a change in the position of the overthrust fault from the "restricted Wissahickon"–Octoraro contact to the "redefined Wissahickon"–limestone contact. In the text of the report, however, Bascom lists six lines of evidence that support an alternate hypothesis to the effect that the albite-chlorite–schist is Ordovician, that the more highly metamorphosed schist to the south is pre-Cambrian, and that the thrust fault belongs several miles south of the valley. The paragraph dealing with the age of the schist of the south valley wall provides an excellent summary of the stratigraphic evidence that the schist is Ordovician:

> The contact of the muscovite schist with the Ordovician limestone of Chester Valley is apparently conformable; this is indicated (a) by a lithologic gradation shown in the micaceous character of the upper limestone beds and the limy character of the lower beds of the mica schist in an apparent syncline at the Howellsville quarries, 1½ miles northeast of Paoli; (b) by the nature of the contact; where this can be observed, as at a small quarry 1½ miles northwest of Gulf Mills, in the Norristown quadrangle, it is either straight and parallel to the stratification of the two formations or there is a refolding of the beds in which both formations participate; (c) by failure to find evidence of the cutting out of limestone beds; south of Gulf Mills, in the Norristown quadrangle, the mica schist shows a synclinal structure with limestone exposed on both limbs and in the axis of the fold to the northeast, beyond Conshohocken; the syncline pitches southwest, and the schist is eroded to the northeast and widens to the southwest until concealed under the thrust fault of pre-Cambrian rocks; (d) by the similarity in chemical constitution respectively of the Martinsburg shale and the conformably (?) underlying limestone of the great valley and the mica schist of the South Valley Hills and underlying limestone of the Chester Valley, which constitutes a presumption in favor of a like age and relationship. Any one of these observations taken singly is capable of another interpretation, but the combined observations seem to the writer to point to similar relations between the mica schist and Cambrian and Ordovician formations and to the Ordovician age of the schist.[22]

In view of the several lines of evidence presented above that the albite-chlorite–schist of the south margin of Chester Valley is Or-

[21] Bliss and Jonas, *op. cit.*

[22] Bascom and Stose, *op. cit.*, p. 5.

dovician, and hence that the more intensely metamorphosed schist into which it grades to the south is also Ordovician, it is desirable to examine briefly the evidence for placing these rocks in the pre-Cambrian. As summarized in a section[23] on "The Age of the Wissahickon" in the McCalls Ferry–Quarryville report, the assignment is based, first, on the placing by Bascom of the "restricted Wissahickon" of the Philadelphia district in the pre-Cambrian; second, on the presence of a discordant contact between the Wissahickon schist and underlying Paleozoic limestones in the Martic Hills region; and finally, on stratigraphic relations observed by Jonas elsewhere in the Piedmont region.

The "Wissahickon mica-gneiss" of the Philadelphia district was first regarded by Bascom as Ordovician, the evidence being presented in a paper[24] that is a classic contribution to our knowledge of the Pennsylvania Piedmont. A note at the end of the paper, however, states that a field conference (held with Messrs. A. Keith, G. O. Smith, and E. B. Mathews to determine the age of the Wissahickon) "while establishing the Ordovician age of the Wissahickon mica schist, place(s) in question the equivalence of the mica gneiss and the mica schist." In the Philadelphia Folio the Wissahickon mica-schist is called the Octoraro (Ordovician) and the Wissahickon mica-gneiss is placed in the pre-Cambrian. The reader must be referred to the text[25] of the Folio for a complete summary of the argument in favor of this change in the age assignment of the mica-gneiss; the most compelling points are, briefly stated, the occurrence of igneous intrusions in the mica-gneiss that are, with the exception of pegmatite, notably absent from known Paleozoic rocks to the northwest, and secondly, certain differences in metamorphic intensity between the Octoraro schist and the Wissahickon mica-gneiss that were regarded as justifying a separation of these rock types into different formations.

It is certainly true that igneous intrusions are more common in the Wissahickon than in Paleozoic rocks farther northwest, but this

[23] Knopf and Jonas, "Geology of the McCalls Ferry–Quarryville District, Pennsylvania," *op. cit.*, pp. 33–35.

[24] Bascom, "Piedmont District of Pennsylvania," *Geol. Soc. Amer. Bull.* 16 (1904), pp. 289–328.

[25] Bascom, "Philadelphia Folio," *op. cit.*, p. 4.

may well be due to difference in geographic position rather than in age. The "igneous unconformity" argument is weakened by the reported presence of gabbro,[26] basalt,[27] granite,[28] serpentine,[29] and many other igneous types intrusive into known Paleozoic rocks in adjacent parts of the Piedmont region. Similarly the difference in the degree of metamorphism between the Octoraro (Ordovician) and the Wissahickon may well be related to such sharp changes in the metamorphic intensity as those discussed by Knopf in the McCalls Ferry–Quarryville Bulletin,[30] or to differences in the original composition of the two rocks. It seems therefore, that the evidence for the pre-Cambrian age of the Wissahickon mica-gneiss in the Philadelphia district is not conclusive.

The discordant relation in the Quarryville district is described as:

> Wissahickon albite-chlorite schist and Peters Creek schist overlying previously folded rocks, the Vintage dolomite, of Lower Cambrian age, and the Conestoga limestone, of Ordovician age.[31]

This outcrop could not be located by the present writer, but in the absence of any statement as to the nature of the contact, it seems entirely possible that the discordance may represent an unconformity rather than a thrust fault. Angular unconformities are widely reported in the Ordovician sequence in the Appalachians; the Cocalico shale, with which the Wissahickon is here tentatively correlated in part, rests unconformably on limestones of lower Trenton and upper Beekmantown age in the immediately adjacent area to the north.[32]

Stratigraphic proof of the pre-Cambrian age of the Glenarm series

[26] C. P. Berkey and M. Rice, "Geology of the West Point Quadrangle, New York," *N. Y. State Mus. Bull. 225–26* (1919).

[27] Stose and Jonas, "Ordovician Shale and Associated Lava in Southeastern Pennsylvania," *Geol. Soc. Amer. Bull. 38* (1927), pp. 505–36.

[28] J. T. Lonsdale, "Post-Cincinnatian Granites of Northeastern Piedmont Virginia," *Jour. Geol.*, Vol. XXXIV (1926), pp. 159–66.

[29] H. H. Hess, "Hydrothermal Metamorphism of an Ultrabasic Intrusive at Schuyler, Virginia," *Amer. Jour. Sci.* Vol. XXVI (1933), pp. 377–408.

[30] Knopf and Jonas, "Geology of the McCalls Ferry–Quarryville District, Pennsylvania," *op. cit.*, p. 119.

[31] *Ibid.*, p. 35. [32] *Ibid.*, p. 58.

(and hence of the Wissahickon) has been described by Jonas in the western Piedmont of Maryland. It is stated that:

> West of Union Bridge basal Cambrian quartz conglomerate unconformably overlies volcanic rocks and near Libertytown rests unconformably on Cockeysville marble. The unconformable relation of basal Cambrian sediments to the volcanic rocks and the Cockeysville marble establishes the pre-Cambrian age of these rocks and of the Glenarm series.[33]

The present writer is not qualified to comment on this evidence, for the critical contact exposures of Cambrian quartzite on Cockeysville marble, although sought for, were not located.

It is not proposed to extend the interpretations presented in this paper beyond the area defined in the introduction, but one interesting and possibly significant relation, described by Jonas in the western Piedmont of Maryland, may be mentioned in passing. The series there exposed includes a marble, interbedded with metamorphosed volcanic rocks including green and blue schists considered to be altered volcanic ash.[34]

The marble-volcanic series is overlain by Wissahickon schist. Jonas states that:

> The period of volcanic activity was preceded and followed by the deposition of calcareous beds of the Cockeysville marble, which in turn was covered by other volcanic outpourings before the deposition of the argillaceous arkose that has been recrystallized with the Wissahickon schist. It seems probable that volcanic activity of an explosive character occurred during the deposition of Wissahickon formation, because in the area south of Union Bridge tuffaceous slates underlie and may be interbedded with the Wissahickon albite schist.[35]

These volcanic rocks are correlated by Jonas with volcanic rocks "known to be pre-Cambrian" in the Pigeon Hills and the Hellam Hills in Pennsylvania.[36]

In a more recent paper dealing with the stratigraphy of the Martinsburg shale, the same author (with Stose) notes the presence of basic lavas in the Ordovician section near Jonestown, Pennsylvania.[37] The lava occurs at the top of the Beekmantown lime-

[33] Jonas, "Pre-Cambrian Rocks of the Western Piedmont of Maryland," *Geol. Soc. Amer. Bull.* 35 (1924), p. 361.

[34] Jonas, *Ibid.*, p. 360. [35] *Ibid.*, pp. 360–61. [36] *Ibid.*, p. 363.

[37] Stose and Jonas; "Ordovician Shale and Associated Lava in Southeastern Pennsylvania," *op. cit.*, pp. 505–36.

stone and is overlain by unaltered red sandstone and shale of the base of the Martinsburg. Certain "hard purple and green shales, which have been observed near the base of the Martinsburg and Cocalico shales at many places are believed to be altered ash, related to the period of vulcanicity observed at Jonestown."[38]

According to the hypothesis here advanced the Cockeysville marble of the western Piedmont of Maryland, with basic lavas interbedded near the top, is of approximately the same age as the Paleozoic limestone of the Jonestown area, which carries similar lavas, similarly placed. The immediately overlying Wissahickon of the Piedmont region, with green and blue tuffaceous slates at the base, is here tentatively correlated with the Martinsburg of the Jonestown area, characterized at the base by green and purple shales "believed to be volcanic ash." The striking similarity of sequence in the two areas supports this correlation.

Summary.—The rectilinear character of the south side of Chester Valley indicates a high-angle contact between the Cambro-Ordovician limestone of the valley and the schist of the south valley side. This and other physiographic and structural relations between the schist and the limestone are more reasonably explained by the hypothesis that the schist rests on the limestone in normal sedimentary succession than by the hypothesis that the schist is a part of a great low-angle–overthrust sheet. The nature of the contact between the two formations indicates that the schist does overlie the limestone in normal sedimentary succession, and is therefore of Paleozoic age.

STRATIGRAPHIC AND STRUCTURAL RELATIONS OF THE SETTERS QUARTZITE AND THE COCKEYSVILLE MARBLE IN THE DOE RUN–AVONDALE REGION

In the area between Doe Run and Avondale, 2 to 8 miles south of Chester Valley, three formations that underlie the Wissahickon schist are exposed in an unroofed anticline. This series includes the Baltimore gneiss at the base, overlain unconformably by the Setters quartzite, the Cockeysville marble, and the Wissahickon schist—the latter three formations representing the lower half of the supposed

[38] Stose and Jonas, *ibid.*, p. 532.

pre-Cambrian Glenarm series. Hawkins, in 1923,[39] questioned the pre-Cambrian age of the Avondale rocks above the Baltimore gneiss, pointing out that this quartzite-marble-schist sequence is essentially the same as the Paleozoic sequence around the Mine Ridge anticline, where the Baltimore gneiss is overlain by Chickies quartzite, Cambro-Ordovician limestones, and Martinsburg shale equivalents. He suggested that the two sequences are actually the same age, basing this conclusion on similarity in the thickness of corresponding members, on the gradation of the Wissahickon schist into the Octoraro schist (considered by him to be Ordovician) and on other structural and stratigraphic relations.

Knopf and Jonas object to Hawkins' interpretation in part on the ground that he compares the sedimentary sequence in the Doe Run–Avondale region

> not with the complete Paleozoic section around the Mine Ridge upland and northward, but with the incomplete section west of Coatesville, where a post-Beekmantown unconformity has removed a considerable thickness of Cambrian and Lower Ordovician beds.[40]

But both the Glenarm and Paleozoic series in this area show marked local cutting out of beds as a result of non-deposition or erosion, and it appears that Hawkins was fully justified in comparing the Doe Run–Avondale section with the *nearest* Paleozoic section (Coatesville is 4 miles from Doe Run) rather than with the lower Paleozoic standard column for the whole region. In any case, thicknesses of incompetent beds, as marble and schist, are not to be regarded as particularly significant in correlation in so intensely deformed a belt as that under consideration.

Hawkins' correlation of the Cockeysville marble of the Avondale district with Cambro-Ordovician limestones near Coatesville is further criticized on stratigraphic grounds involving the positions of certain unconformities and overlap relationships. Space does not permit discussion of these questions of detailed stratigraphy here; indeed, so complete is the recrystallization of the marble in the

[39] A. C. Hawkins, "Alternative Interpretation of some Crystalline Schists in Southeastern Pennsylvania," *Amer. Jour. Sci.*, 5th ser., Vol. VII (1924), pp. 355–64.

[40] Knopf and Jonas, "Geology of the McCalls Ferry-Quarryville District, Pennsylvania," *op. cit.*, p. 21.

Avondale region that it seems improbable that it can be subdivided to the extent possible in the Paleozoic limestones to the north. In any case the same stratigraphic arguments may be urged with much greater force against the theory that the Glenarm series as a whole differs in age from the adjacent Cambro-Ordovician series. Supposedly pre-Cambrian Glenarm rocks, resting on Baltimore gneiss, outcrop most extensively south of the supposed Martic overthrust fault, and total over 10,000 feet in thickness. Yet, although the basal Cambrian contact has been studied over large areas north of the indicated position of the fault, *no* Glenarm rocks have been positively identified between the Baltimore and the overlying Cambrian quartzite. It is evident, therefore that the theory that the Glenarm series is pre-Cambrian requires either non-deposition of Glenarm sediments in the northwestern part of the area during a period when at least 10,000 feet of Glenarm sediments were being deposited to the southeast, or, as seems equally improbable, uplift and complete pre-Cambrian erosion of Glenarm rocks in the northwestern part while at least 10,000 feet were preserved to the southeast.

A third objection to Hawkins' interpretation is stated as follows:

> the aggregate thickness of the Wissahickon formation and the Peters Creek schist as exposed along Susquehanna River must be 8,000 to 10,000 feet, whereas the total thickness of all the Ordovician argillaceous sediments exposed south of Blue Mountain is probably of an order of magnitude of only 3,000 feet.[41]

But according to Behre, a recognized authority on Pennsylvania slates, the thickness of the Martinsburg in Northhampton County, Pennsylvania, is approximately 11,000 feet.[42] Here again, it must be emphasized that little significance is attached to comparison of thicknesses of shales and schists.

In the earlier report on the Doe Run–Avondale district Knopf and Jonas correlated the quartzite and marble of the Avondale series with those of the Paleozoic sequence of the south limb of the Mine Ridge anticline.

> The rock (quartzite) south of the North Valley Hills does not contain Scolithus linearis, but from its lithologic character and structural relations may be

[41] Knopf and Jonas, *ibid.*, p. 23.

[42] C. H. Behre, Jr., "Slate in Pennsylvania," *Pa. Geol. Surv. Bull. M 16* (1933), 4th ser., p. 137.

correlated with the Chickies quartzite of the North Valley Hills. . . . . That the limestone of the Doe Run and Avandale region is to be correlated with the Cambrian and Ordovician Shenandoah limestone of Montgomery and Delaware counties is proved by the stratigraphic continuity of the limestone in Chester Valley, by marked lithologic resemblance between the limestone of the Doe Run and Avondale region and the rock in Chester Valley, and by the position of the limestone south of Chester Valley in approximately the same line of strike as the narrow belt of Shenandoah limestone in Cream Valley near Conshohocken, Pa.[43]

To this evidence it may be added that both the Chickies and the Setters are characterized by stretched and broken black tourmaline crystals and that Scolithus has been reported in the Setters.[44]

Since the Wissahickon was regarded as pre-Cambrian in this first Avondale report a thrust fault was drawn between it and the quartzite and limestone (then considered Paleozoic) on which it rests. The Avondale marble-quartzite area was thus considered a window, and the existence of the encircling thrust fault was justified as follows:

In order to explain the presence of a pre-Cambrian gneiss (the Wissahickon) in immediate contact with the Cambrian and Ordovician limestone and the Ordovician schist it is necessary to assume the existence of a deep overthrust fault, the origin and nature of which have been described and which has brought the older formation along a gentle hade to lie upon the younger Paleozoic rocks. The resulting discordance has been obscured by a subsequent diastrophic movement that has folded and faulted the thrust plane as if it were a plane of stratification.[45]

In an abandoned quarry 1 mile northwest of Avondale, Cockeysville marble and Wissahickon schist are exposed and the junction plane appears to be a gradational sedimentary contact, not a fault plane. This contact, and others of similar nature in the immediate vicinity, indicate that whatever the absolute age of the two formations may be, their relative age is nearly the same. Evidence has been adduced above that suggests that the marble is Cambro-Ordovician, and probably equivalent to the Chester Valley lime-

[43] Bliss and Jonas, "Relation of the Wissahickon Mica Gneiss to the Shenandoah Limestone and the Octoraro Schist of the Doe Run and Avondale Region, Chester County, Pennsylvania," *op. cit.*, pp. 17, 19. This paragraph does not express the present interpretation of these workers.

[44] H. D. Rogers, P. Frazer, and C. E. Hall, "The Geology of Chester County," *Second Geol. Surv. Pa. Bull. C4* (1883), p. 333.

[45] Bliss and Jonas, "Relation of the Wissahickon Mica schist to the Shenandoah Limestone, etc.," *op. cit.*, p. 33.

stone. It was shown in the last section that there is good ground for believing the Wissahickon schist to be Ordovician. The contact is strikingly similar to the contact between Cambro-Ordovician limestone and Wissahickon schist at the south side of Chester Valley

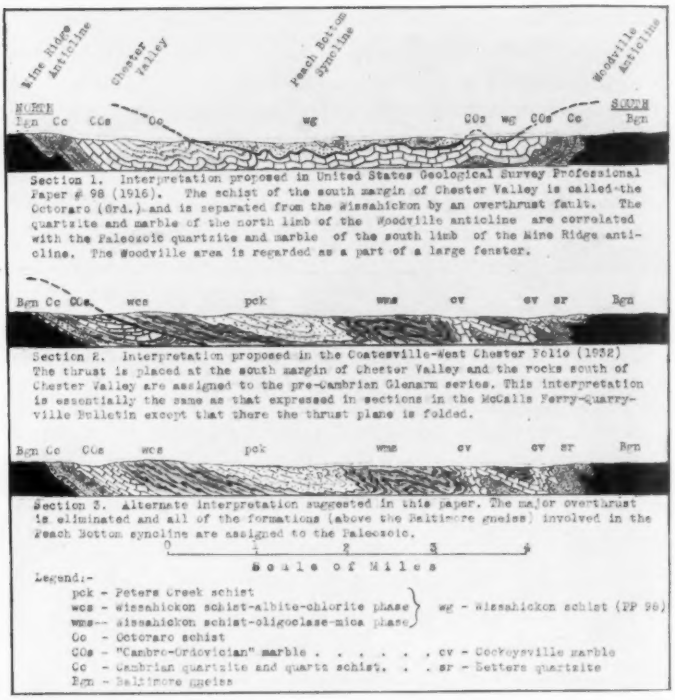
Mine Ridge Anticline
Chester Valley
Peach Bottom syncline
Woodville Anticline
NORTH
Bgn Cc COs Cc wg COs wg COs Cc SOUTH Bgn
Section 1. Interpretation proposed in United States Geological Survey Professional Paper # 98 (1916). The schist of the south margin of Chester Valley is called the Octoraro (Ord.) and is separated from the Wissahickon by an overthrust fault. The quartzite and marble of the north limb of the Woodville anticline are correlated with the Paleozoic quartzite and marble of the south limb of the Mine Ridge anticline. The Woodville area is regarded as a part of a large fenster.
Bgn Cc COs wcs pck wms cv cv sr Bgn
Section 2. Interpretation proposed in the Coatesville-West Chester Folio (1932) The thrust is placed at the south margin of Chester Valley and the rocks south of Chester Valley are assigned to the pre-Cambrian Glenarm series. This interpretation is essentially the same as that expressed in sections in the McCalls Ferry-Quarryville Bulletin except that there the thrust plane is folded.
Bgn Cc COs wcs pck wms cv cv sr Bgn
Section 3. Alternate interpretation suggested in this paper. The major overthrust is eliminated and all of the formations (above the Baltimore gneiss) involved in the Peach Bottom syncline are assigned to the Paleozoic.
0 1 2 3 4
Scale of Miles
Legend:-
pck - Peters Creek schist
wcs - Wissahickon schist-albite-chlorite phase
wms - Wissahickon schist-oligoclase-mica phase
wg - Wissahickon schist (PP 98)
Cc - Octoraro schist
COs - "Cambro-Ordovician" marble . . . . . cv - Cockeysville marble
Cc - Cambrian quartzite and quartz schist. . . sr - Setters quartzite
Bgn - Baltimore gneiss

FIG. 4.—Alternate interpretations of the structure between the Mine Ridge anticline and the Woodville-Avondale compound anticline. Location and direction of sections is shown by line *A-A* in Figure 2.

near Downingtown; the Cockeysville and Wissahickon of the Avondale exposures are similar in lithology to the Cambro-Ordovician limestone and the Wissahickon, respectively, of the Downingtown cut.

In remapping the Avondale district for the Coatesville-West

Chester Folio, Bascom removed the thrust fault[46] from around the outcrop area of quartzite and marble formerly interpreted as a window, and represented the schist as resting on the limestone in normal sedimentary succession. Since, however, the Wissahickon was retained in the pre-Cambrian, the superposition of schist upon the limestone necessitated placing the latter formation and the associated quartzite also in the pre-Cambrian.

Thus it appears that here again, as in the Chester Valley area, the apparent contradictions and reversals of structural interpretation are introduced (and the overthrust is necessitated) by the assignment of the Wissahickon to the pre-Cambrian. The evidence for this age assignment has been reviewed briefly, and found not to be conclusive. The structural interpretation advanced here as an alternate to the two already mentioned is as follows: In simplest terms, the major structural elements of the region are the Mine Ridge anticline, the Avondale (Woodville) anticline, and an intervening syncline. The Peters Creek schist outcrops in the center of the syncline, the oligoclase-mica facies of the Wissahickon, the Cockeysville marble, and the Setters quartzite of the south limb (and the Avondale anticline) being equivalent, respectively, to the albite-chlorite facies of the Wissahickon, the Cambro-Ordovician limestone, and the Chickies quartzite, of the north limb (and the Mine Ridge anticline) (Fig. 4.). This interpretation is similar to that proposed by Bascom, and the synclinal axis coincides with that of the Peach Bottom syncline of the Coatesville–West Chester Folio, but it differs from Bascom's interpretation in that all of the formations involved, (except the Baltimore gneiss) are considered Paleozoic, and in that no major overthrust is placed between the albite-chlorite facies of the Wissahickon schist and the Chester Valley limestone.

STRATIGRAPHIC RELATIONS OF THE PEACH BOTTOM SLATE

The Cardiff conglomerate and the Peach Bottom slate are the youngest members of the Glenarm series. Both are preserved in a downfold, the Peach Bottom syncline, which is continuous from the

[46] Bascom and Stose, "Coatesville–West Chester Folio," *loc. cit.* In an earlier paper Knopf and Jonas imply that they no longer consider the Avondale area to be a window. "Stratigraphy of the Crystalline Schists of Pennsylvania and Maryland," *Amer. Jour. Sci.*, 5th Ser., Vol. V, (1923), p. 42.

Coatesville–West Chester region southwestward to the Susquehanna River and beyond, the type locality of the slate being near Peach Bottom village, in the Susquehanna gorge. Since the Cardiff and the Peach Bottom are not in contact with known Paleozoic rocks in the area under consideration the question of the age of these formations must depend, first, on paleontologic evidence if any is available, and second, on their relation to known Paleozoic beds elsewhere in the Piedmont region.

In 1876 certain supposed fossils were obtained from the dump piles of the Peach Bottom roofing slate quarries and later sent to James Hall, who communicated his findings to Persifor Frazer in 1883. Hall's letter is quoted, in part, by Knopf and Jonas in discussion[47] of the age of the Peach Bottom, but the last paragraphs of the letter, in which Hall expresses his considered judgment of the fossil assemblage as a whole, are not quoted. Hall states:

> Although I can find nothing which will give positive information of the actual horizon of these slates in the geologic series, *there can be no doubt but that they occupy a place either in the Hudson River or Quebec series*, and not unlikely in the former.
>
> I regret that I am unable to give more positive information, but the facts within my knowledge do not warrant a positive expression.[48]

Two questions may be raised as to the validity of this evidence; first, as to whether the slabs examined by Hall actually came from the Peach Bottom locality, and second, as to the correctness of Hall's interpretation. The fact that Persifor Frazer and other Pennsylvania geologists by whom the rocks were sent to Hall applied his age determination to the Peach Bottom slates gives reasonable assurance as to the authenticity of the material. Hall's reputation and the fact that Leo Lesquereux reported independently that certain of the markings represented *Buthotrephis flexuosa* are the best available answers to the question as to the interpretation of age.

Knopf and Jonas, however, state that:

> the most careful and persistent search through the Peach Bottom slates by numerous geologists has failed to reveal any further fossil markings that would confirm the reported find of graptolites.[49]

[47] "Geology of the McCalls Ferry–Quarryville District, Pennsylvania," *op. cit.*, p. 40.

[48] P. Frazer, Jr., "The Peach Bottom Slates of Southeastern York and Southern Lancaster Counties," *Trans. Amer. Inst. Min. Met. Eng.*, Vol. XII (1884), p. 358 (italics not in original).

[49] "Geology of the McCalls Ferry–Quarryville District, Pennsylvania, *op cit.*, p. 41.

Dr. H. H. Hess and the writer were unable, in a brief examination of the region, to locate outcrops favorable (those in which slaty cleavage parallels the original sedimentary bedding) for fossil collecting. A search of several quarry dumps where this favorable condition did not exist failed to disclose undoubted fossil forms. For this reason the Frazer-Hall determination of the age of the Peach Bottom as Ordovician is presented here on its own merits.

The Glenarm series has been traced southward from Pennsylvania through Maryland into Virginia. In a recent paper[50] Jonas describes certain Virginia slates, associated with Glenarm rocks, which have been shown to be Ordovician by fossil evidence. It is then stated that

> . . . . the Arvonia and Quantico slate lie in the southwestward extension of the Peters Creek syncline. In Maryland and Pennsylvania similar slates, the Peach Bottom, which occur in the same syncline, have yielded no fossils and have been placed in the pre-Cambrian in the absence of positive evidence to the contrary. The presence of slate of known Ordovician age in the same syncline in Virginia is a matter of interest.[51]

The present writer has not examined the Arvonia and Quantico localities, but the reported presence of these Ordovician slates in line of strike with the Peach Bottom seems to afford strong support to the assignment of the Peach Bottom to the Ordovician, and to present a formidable difficulty to the theory that it is pre-Cambrian.

CONCLUSION

Evidence presented in this paper supports the hypothesis that the Glenarm series of metamorphic rocks of the Pennsylvania Piedmont region is Paleozoic rather than pre-Cambrian in age. If this theory is correct it is not necessary to suppose that a large part of the Piedmont region has been moved northwestward some 20 miles along the Martic overthrust fault. The primary importance (in so far as lower Paleozoic stratigraphic studies, petrologic studies of Piedmont igneous and metamorphic history, and other lines of geologic investigtion are concerned) of definitely determining the age of the Glenarm rocks can hardly be over-emphasized. It is the purpose of

[50] "Geologic Reconnaisance in the Piedmont of Virginia," *op. cit.*, pp. 837–46.

[51] *Ibid.*, p. 842.

this paper to reopen a problem that, in the area under consideration[52] has apparently been considered closed.

ACKNOWLEDGMENTS.—In the course of a field excursion in southeastern Pennsylvania in 1932 Professor Douglas Johnson first called attention to the fact that the straight south margin of Chester Valley is incompatible with the theory that it was determined by erosion along a low-angle thrust fault. When the writer discovered additional evidence of the non-existence of such a fault Professor Johnson encouraged prosecution of the study and offered helpful criticism. The paper was first presented before the Seminar in Geomorphology at Columbia University during the autumn of 1933. The text has been read by Professors Agar, Balk, Berkey, Colony, and Johnson, and by Dr. H. H. Hess, but the interpretations advanced do not necessarily reflect the views of these geologists. Particular thanks are due to Dr. H. H. Hess for aid in the field work and in the preparation of the manuscript. The tentative conclusions proposed here differ in important respects from those held by Mrs. Eleanora Bliss Knopf and Miss Anna I. Jonas. For this reason the writer wishes especially to express his sincere appreciation of the truly great contributions made by these workers to our knowledge of Piedmont geology.

[52] The Manhattan schist and associated metamorphic rocks of southeastern New York state are commonly correlated with the Wissahickon schist and related Glenarm rocks of the Pennsylvania Piedmont. Dr. Robert Balk, on the basis of extended and most detailed field studies, has recently presented strong evidence that the Manhattan series is Paleozoic.

GLACIERS OF THE GRAND TETON NATIONAL PARK OF WYOMING[1]

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ABSTRACT

Explorations made during the summers of 1926–31 indicated the existence of at least seven small glaciers in the Teton Range of northwestern Wyoming. These occur on the east slope of the range, within the rugged area now comprising Grand Teton National Park, at altitudes between 10,000 and 11,500 feet. Herein is given the setting, description, and history of these glaciers.

THE TETON RANGE

From its northern extremity in southwestern Yellowstone Park the Teton Range extends south-southwestward with unbroken continuity for 40 miles before losing its identity in the Teton Pass Mountains, which serve to link it with the transverse Snake River Range. Throughout this course the range forms a barrier between two adjacent intermontane basins—Jackson Hole, Wyoming, on the east, and Teton Basin, Idaho, on the west. Springing with wall-like abruptness from the floors of these basins the snow-clad range is surpassingly majestic. Its higher peaks attain altitudes of 12,000 to nearly 14,000 feet, towering a mile or more above the plains of Jackson Hole (the latter being 6,000–7,000 feet above sea level).

The range[2] is of the fault-block type, the pre-Cambrian complex and overlying sedimentaries being fractured along a north-south line, uplifted, and tilted westward. The principal fault, whose displacement may be as great as 15,000 feet, for the most part follows the straight east border of the range; hence the absence of foothills on this side. The narrow east face of the range exposes the pre-Cam-

[1] Report prepared for the Division of Research and Education, National Park Service.

[2] Orestes St. John, "Report of Teton Division," *U.S. Geol. and Geog. Surv. of the Territories, 11th Ann. Rept.* (1877), pp. 321–508; Eliot Blackwelder, "Post-Cretaceous History of the Mountains of Central Western Wyoming," *Jour. Geol.*, Vol. XXIII (1915), pp. 97–117, 193–217, 307–40; Blackwelder, "A Reconnaissance of the Phosphate Deposits in Western Wyoming," *U.S. Geol. Surv. Bull., 470* (1911), pp. 452–83; M. R. Campbell and others, Geologic Map of Wyoming (U.S. Geol. Surv., 1925).

brian core, and these ancient rocks are also laid bare throughout extensive portions of the broad, gently inclined west slope from which the Paleozoic and Mesozoic strata have been stripped away.

Evidences of glaciation occur on every hand, most strikingly in the crystalline pre-Cambrian areas, where the steeply pitching gneisses and schists lent themselves with peculiar effectiveness to glacial quarrying. Capacious, broad-floored canyons terminating in typical cirques, intervening knife-like aiguilles which lead up to a score or more peaks of the "horn" type—these bespeak for the Teton Range a glacial experience of exceptional intensity and lend it a severe grandeur probably unparalleled among western ranges of the United States. The glacial sculpturing is seen at its best in the 25-mile section of the east slope between Granite and Moose canyons, the area which in 1929 was designated a national park and to which attention is directed in this report (Fig. 1).

HISTORY OF EXPLORATION

It might be expected that remnants of the ice bodies which in Pleistocene time so profoundly modified the topography of these mountains should still exist, but the literature relating to this question is contradictory. On the Grand Teton Quadrangle (surveyed in 1898–99), for example, no glaciers are indicated; and in the notes descriptive of this map which were published by the United States Geological Survey in 1914 it is stated that, "owing to the small snowfall it [the Teton Range] carries no glaciers, although there are indications here and there of the presence of small glaciers in prehistoric time."[3] On the other hand, accounts published by mountaineers and others who have made casual observations in the range imply that glaciers are not only present but numerous. By this group the term "glacier" is all too often used indiscriminately for almost any field of snow or ice.

F. V. Hayden was the first geologist to make observations in this region; however, in crossing Teton Pass with General W. F. Raynolds in June, 1860, he had no opportunity to see, except from a distance, that section of the range in which are found the glaciers, and

[3] "Set of Fifty Atlas Sheets Showing Physiographic Types and Features of Interest to Engineers," *U.S. Geol. Surv. Circ., 9–323 y*, p. 3.

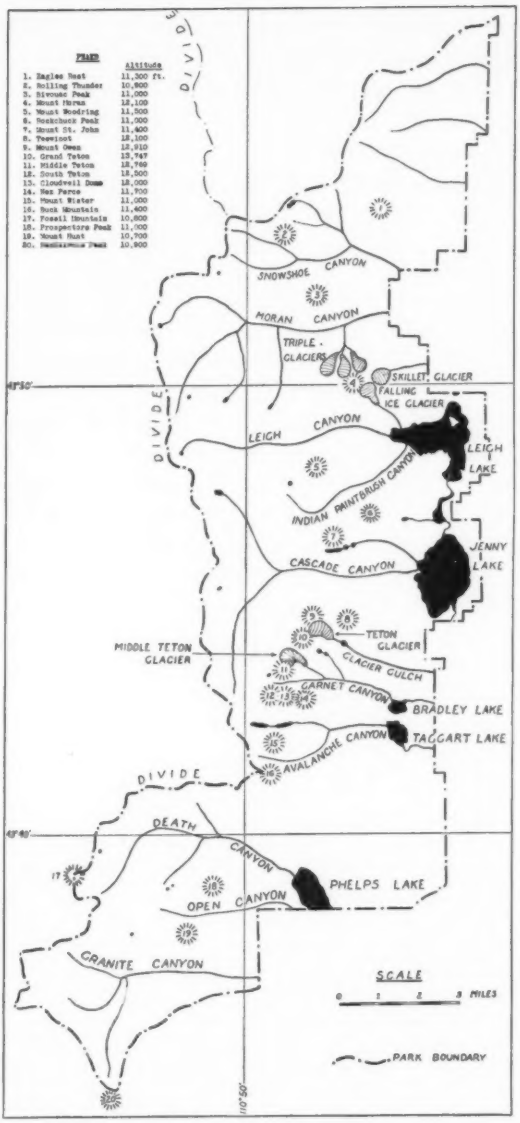

FIG. 1.—Sketch map, Grand Teton National Park of Wyoming (showing the location of the glaciers described in this paper).

hence these remained unreported.[4] In 1872, F. H. Bradley of the Hayden Survey invaded the higher Tetons and must have reached points[5] affording views of Middle Teton Glacier; the latter, however, escaped discovery. Subsequently from the Jackson Hole side Bradley made the less pardonable error of concluding, after very limited observation, that "the water of these mountain streams is so pure as to make it certain that not the least glacial erosion is now going on at any point on the range."[6] Orestes St. John, of the Hayden Survey of 1877, was conservative to the extent of being noncommittal respecting the importance of past glaciation.[7]

Recognition of the Teton glaciers came in the summer of 1878, when a party of the Hayden Survey (including Dr. Hayden himself), then engaged in its final season of field work, in traveling from the Wind River Mountains to Yellowstone Park skirted the east shore of Jackson Lake. Examinations made with aid of field glasses convinced these observers of the glacial character of some of the ice bodies on Mount Moran; and in the subsequent publications of W. H. Holmes and James Eccles this conclusion was set forth.[8] In 1885 brief notes by Holmes were incorporated in I. C. Russells' paper on "Existing Glaciers of the United States,"[9] together with an illustration of Mount Moran showing Skillet Glacier.

In succeeding decades geologists had opportunity to view the Teton glaciers only from remote points, and consequently only a few brief references made their way into the geologic literature.[10] So

[4] F. V. Hayden, *Geological Report of the Exploration of the Yellowstone and Missouri Rivers* (Washington, 1869).

[5] F. V. Hayden, *U.S. Geol. Surv. of the Territories, 6th Ann. Rept.* (1872), p. 220.

[6] *Ibid.*, p. 264; F. H. Bradley, "Explorations of 1872: U.S. Geological Survey of the Territories, under F. V. Hayden; Snake River Division," *Amer. Jour. Sci. and Arts*, 3d Ser., Vol. VI (September, 1873), pp. 194–207.

[7] *Op. cit.*, pp. 419–20.

[8] W. H. Holmes, "Report on the Geology of the Yellowstone National Park" in *U.S. Geol. and Geog. Surv. of the Territories, 12th Ann. Rept.* (1878), by F. V. Hayden, Vol. II, p. 52, James Eccles, "The Rocky Mountain Region of Wyoming and Idaho," *Alpine Jour.* (London), Vol. IX (1878–79), p. 251.

[9] *U.S. Geol. Surv., 5th Ann. Rept.* (1885), pp. 346–47.

[10] Arnold Hague, "Geology of Yellowstone National Park," *U.S. Geol. Surv. Folio 30* (1896), p. 3; C. K. Wentworth and D. M. Delo, "Dinwoody Glaciers, Wind River Mountains, Wyoming; with a Brief Survey of Existing Glaciers in the United States," *Geol. Soc. Amer. Bull. 42* (1931), p. 610.

far as known, no studies close at hand were made prior to those herein summarized. The latter extended through the summers of 1926–31, inclusive, during which period the alpine portion of the range was systematically explored. Traverses of most of the canyons and ascents of the peaks made possible a scrutiny of those remote recesses in which glaciers might conceivably lie; and, so far as possible, visits were made to ice fields that gave evidence of glacial character.[11]

DISTRIBUTION OF THE TETON GLACIERS

The glaciers herewith reported (Fig. 1)—seven in number—occur on the east slope of the range at elevations ranging from 10,000 feet to 11,500 feet; and all can be seen, though not to equal advantage, from points in Jackson Hole. Like most glaciers in the Rocky Mountains of this country, they are small, in no case attaining a mile in length. From the glaciers themselves, or equivalent altitudes elsewhere in the range, one obtains magnificent views southeastward across Jackson Hole to the mountains beyond, where, on clear days, the far more imposing Wind River glaciers, 70 miles away, may be easily distinguished.[12]

Here, as in some other localities in the Rockies, the restriction of existing glaciers to the slopes east of the divide is undoubtedly in a measure due to the tendency of the light, dry winter snow to drift with the prevailing southwest winds and collect in appropriate catchment basins on the lee side of the range. More significant in explaining the distribution of the existing glaciers, however, is their relationship to the major peaks. With but one exception (Falling Ice Glacier) the Teton glaciers face north or east and lie in the shadow of major peaks, whose sheer walls, rising to the south and west and towering hundreds, or even thousands, of feet above them, have furnished the protection without which they would long since have dwindled to snow fields. Those peaks which do not now carry a glacier on their north sides must, in most cases, have done so up to

[11] Previous publications by the writer relating to the Teton glaciers and the glaciation of the Teton Range are as follows: *Glacial Features of Jackson Hole, Wyoming*, "Augustana Library Publ.," Vol. XIII (Rock Island, Illinois: Augustana College, 1930; 130 pp.); "The Formation of Glacial Tables," *Jour. Geol.*, Vol. XLI (1933), pp. 642–46; "The Migration of Superglacial Boulders," *ibid.*, pp. 737–47.

[12] Wentworth and Delo, *op. cit.*, pp. 611–20.

relatively recent times, as indicated by their persistent fields of névé, some of which retain glacial characters.

The position of the great Teton peaks is, therefore, an important factor in the localizing of glaciation. In an earlier erosional stage of

FIG. 2.—Middle Teton Glacier, in 1926. In the background is the south head; in the foreground, the main body of the glacier. The crevasses of these divisions trend at right angles to each other, and where the two systems meet there is off-setting. In the upper right-hand corner are the basal cliffs of the Middle Teton.

the range the peaks coincided, at least approximately, with the divide. Owing to the exceedingly high gradients of the east-flowing streams, as compared with those of the west slope, and the resulting more rapid retreat of their canyon heads, the divide migrated westward until remnants of it became detached and were left standing above the widening east slope. These remnants, greatly modified by

glacial sculpturing, constitute the Teton peaks. Thus is explained the fact that the Middle Teton, the Grand Teton, and Mount Moran, and the glaciers which they support, are found not along the divide but respectively 2 miles, $2\frac{1}{2}$ miles, and 5 miles to the east of it.

Any investigation of small glaciers such as those occurring in the Tetons and elsewhere in the Rockies of the United States is likely to be attended with certain difficulties. At times, particularly in spring and early summer, the glaciers may be snow-laden; at other times, later in the season, partially covered with rock débris. In either case extensive areas of ice may be hidden. There is also the vexing problem of maintaining a distinction which nature itself does not make: of drawing a line between true glaciers and the snow fields into which, in their decline, glaciers eventually and gradually pass. Among geologists this distinction has furnished occasion for divergence of opinion.[13] In the present report consideration is limited to ice bodies whose glacial character would, in the writer's opinion, be conceded without question by most glacialists. "Vestigial glaciers" which still exhibit more or less glacial character are abundant in the range.[14]

MIDDLE TETON GLACIER

Middle Teton Glacier (Fig. 2) occupies the upper end of the north fork of Garnet Canyon between the Middle and Grand Tetons, just east of the saddle connecting the two peaks. The glacier proper consists of a narrow ice tongue about half the width of the canyon floor. It lies close to the south margin of the canyon, hugging the basal

[13] E.g., see: S. F. Emmons, "On Glaciers in the Rocky Mountains," *Proc. Colo. Scientific Soc.*, Vol. II (1887), pp. 211–27; Junius Henderson, *Extinct and Existing Glaciers of Colorado*, "Univ. of Colo. Studies," Vol. VIII (1910), pp. 33–76.

[14] One such occurs on the northeast slope of Mount Owen and was perhaps responsible for the name "Glacier Canyon" formerly applied to Cascade Canyon. Another lies within the forks of this canyon, on the east face of the Table Mountain mass; this has not been examined at close range by the writer but may prove to be a true glacier despite its small size, since the stream descending from it is milky. Ice bodies which similarly deserve closer examination occur on the north face of the ridges which extend westward from Mount Moran and Bivouac Peak. One "vestigial glacier" which discharges into a lake lies at the southwest foot of the Middle Teton; another, on the northeast slope of Mount Moran; still another, at the head of Avalanche Canyon, between Mount Wister and the divide. Other bodies more or less glacier-like have been noted in the heads of the canyons on the west slope. Future observations will determine whether, among such remnants, there may not still be some that should be ranked as true glaciers.

cliffs of the Middle Teton, which have made possible its persistence under conditions otherwise unfavorable.

A well-developed lateral moraine, sinuous and with sharp crest, follows the north margin of the glacier. The south lateral moraine is confined to the lower portion of the tongue, beginning where the ice becomes free of the Middle Teton cliffs; consequently it is only one-third as long as the north lateral and is likewise less sharply defined. The two ridges meet in front of the narrow extremity of the glacier in an inconspicuous terminal moraine.

The glacier proper is fed by two short tributary branches, one from the south (Fig. 2) and one from the west, the two being separated by the west (main) peak of the Middle Teton. On each, bergschrund and névé field are distinct. The west branch extends with gentle slope almost to the saddle, being encircled by cirque cliffs only 50–100 feet high. The south branch comes down steeply from a re-entrant between the east and west peaks of the Middle Teton. Both branches, as well as the glacier proper below the point of their union, are well crevassed. Being principally of the transverse type, the crevasses of the south branch extend very nearly at right angles to those of the west branch. As shown in Figure 2, the lowermost crevasses of the south branch cut and offset a group of small crevasses which have formed at right angles to them.

The amount of surface débris on the lower portion of this glacier is very great, in 1926 and 1931 constituting an almost continuous mantle, derived largely from the north cliffs of the Middle Teton.

In the latest glacial stage, the Pinedale (correlated with the Wisconsin stage of the Middle West), Middle Teton Glacier occupied the full length of Garnet Canyon and beyond its mouth expanded on the floor of Jackson Hole, where it built the terminal moraine which encloses Bradley Lake. At its greatest extent this glacier was 5 miles long. Its most important affluent lay between the South and Middle Tetons, at the head of the south fork of Garnet Canyon, a cirque which now contains no glacial remnant, though persistent snow fields still lie on the north slopes of Cloudveil Dome and the South Teton.

Garnet Canyon is relatively easy to ascend, and reaching the glacier from the head of the north fork involves only the simple ex-

pedient of crossing the north lateral moraine, which nowhere exceeds 75 feet in height. The glacier itself may be explored with ease and safety.

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FIG. 3.—Teton Glacier, photographed in 1928 from near Amphitheater Lake. East ridge of Grand Teton at left; Mount Owen in the center, behind the glacier. Only the terminal lobe of the glacier is clearly disclosed in this view, most of the main field and all of the head lying behind the east ridge of the Grand Teton. Note the prominent moraine that encircles the ice terminus, and the surface débris of the glacier.

TETON GLACIER

Teton Glacier (Figs. 3–5) which occupies the great east-facing cirque between the two highest peaks in the range, the Grand Teton

and Mount Owen, is the only Teton glacier which has been made accessible by trail, being now readily attained by the trail to Amphitheater Lake. The ridge immediately north of Amphitheater Lake (which has an altitude of 9,700 feet) affords a panoramic view (Fig. 3) of the cirque, the glacial snout protruding from it, and the crescentic moraine which the latter has built out into the canyon; and

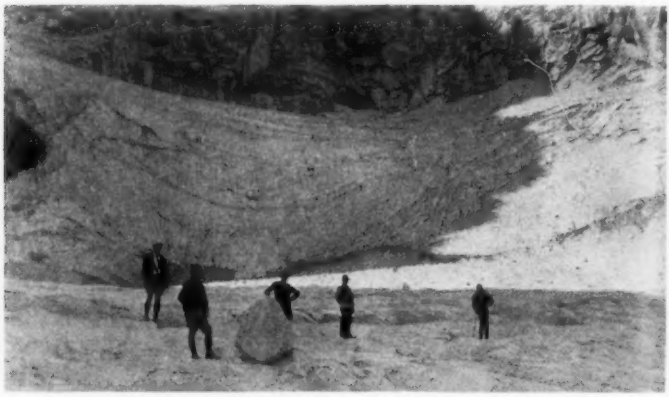

FIG. 4.—Upper portion of Teton Glacier, photographed in 1932. The men are standing near the west edge of the nearly featureless central field; in the background, sloping steeply upward toward the cirque wall, is the crevassed head of the glacier. (Photo by A. W. Gabby.)

one can cross without difficulty from the lake to the ice, a half-mile scramble.

The glacier is about three-fourths of a mile in length and, except in the lower portion, occupies the full width of its cirque. Its course is not due eastward, for it bends somewhat toward the right, curving around the bulging north base of the Grand Teton as does Middle Teton Glacier with reference to its protecting peak. The cirque is one of the most impressive in the range, with walls, exceptional alike for their tremendous height and sheerness, which effectually shelter it. Significantly, the most imposing of these walls, the great north precipice of the Grand Teton, is to the south. Its height is about 3,500 feet.

Three divisions may be recognized in Teton Glacier: the uppermost portion, or "head"; a central field; and a terminal lobe. The first-mentioned is a steep, relatively narrow part of the glacier which descends from the arête forming the west cirque wall (Fig. 4). It is strikingly fissured, toward the end of each summer exhibiting scores

FIG. 5.—Crescentic moraine encircling terminal lobe of Teton Glacier, photographed in 1929. The same moraine, seen from the outer side, appears in Figure 3.

of huge transverse crevasses distributed between the bergschrund and its lower margin.

The much more extensive central area also appears, in part, in Figure 4. This division extends eastward in a continuous, broad field which is commonly snow-covered and which, unlike the portions both above and below it, rarely becomes crevassed except near the north margin. This central area may be described as a very broad shallow trough inclined gently toward the east. Its slopes are steep only where marginal snow fields accumulate against the north and south cirque walls.

Eastward the central field narrows and, with a marked steepening,

gives off the lowermost division, a lobe-shaped terminal extension, easily distinguished by its steeper slope (in places approaching 20°) and its convex surface (Fig. 5). Each summer the snow cover of this portion melts away, exposing the blue ice beneath and scores of crevasses and moulins. The crevassing of the terminal lobe is probably due to several factors: to tension set up near its head, where the gradient steepens; to the fact that this portion of the glacier is thrust beyond the confining walls of the cirque proper into the broader canyon, where lateral spreading takes place; and to the fact that the bend of the glacier toward the right is here most emphasized, leading to tension and consequent fracturing along the north margin.

A finely developed moraine forms a great loop around the margin of the terminal lobe, its sharp even crest (Figs. 3 and 5) being interrupted only where a glacial stream makes its escape. The inner face of the moraine is, ordinarily, only 40–60 feet high; the outer face is everywhere much higher (Fig. 3), extending steeply down into the canyon for hundreds of feet.

Innumerable surface rills from the central field come together on the terminal lobe to form streams, some of which plunge into crevasses and moulins, others of which continue over the surface to the end of the glacier where, converging, they give birth to Glacier Creek. The latter breaks through the notch at the lowest point of the moraine and cascades down its steep outer face to the floor of the canyon. Half a mile lower it enters Delta Lake, a settling basin wherein is deposited much of its suspended rock-flour. That not all of the sediment brought into Delta Lake is deposited there is evident from the fact that Glacier Creek is ordinarily milky all the way to the foot of the range, and beyond.

At its fullest extent in Pinedale time Teton Glacier was $3\frac{1}{2}$ miles long, reaching down to the base of the range and depositing a terminal moraine on the floor of Jackson Hole. When, later, the ice terminus retreated, the basin within this moraine became available for a lake, subsequently destroyed by filling. The swampy meadow inside the moraine indicates the form and extent of this ancient lake. No doubt here, as in the case of Delta Lake, rock-flour from the glacier constitutes most of the filling, and it seems likely that this lake became extinct before Teton Glacier had retreated far enough

to expose the basin of Delta Lake, for otherwise the latter, being farther upstream, would have been the first to be eliminated.

FALLING ICE GLACIER

From the southeast slope of Mount Moran rise two pointed "horns," each about a thousand feet high; between and behind these,

FIG. 6.—Upper portion of Falling Ice Glacier, behind its two protecting "horns." The terminal ice tongue extends down between these horns in the direction of Leigh Lake, below. In the view appear some of the crevasses formed where the glacier steepens toward the head of the ice tongue. Photograph made from summit of Mount Moran in 1930.

countersunk in the mountain, is the cirque of Falling Ice Glacier. This unusual little glacier has been enabled to persist on the sunny face of the mountain by reason of the depth of its cirque and the protection received from the huge horns at its portals (Fig. 6). Despite the southeast exposure, only for a limited portion of each day does direct sunlight have access to the surface of the ice. In Pinedale time, however, Falling Ice Glacier reached out beyond the horns and far down the slope of Mount Moran to join the large glacier which then emerged from the mouth of Leigh Canyon.

This glacier is gourd-shaped, consisting of the nearly circular upper portion which floors the cirque and a long tongue of ice which is sent off southeastward into the narrow cleft between the two horns (Fig. 7). Both above and below the glacier the great diorite dike of Mount Moran outcrops, as shown in the illustration. The upper portion of the glacier has a basin-like surface, sloping in toward the center from all sides except the southeast, where, between the horns, the

FIG. 7.—Terminal precipice of Falling Ice Glacier, between the "horns" of Mount Moran, as it appeared in 1931. The diorite dike passes beneath the upper portion of the glacier, which lies in the cirque behind the two horns. Note the absence of morainal accumulations, due to the steepness of the slope down which the ice tongue has been thrust. As the camera had to be pointed upward about 30° to obtain this photograph, the true steepness of the slope is not apparent.

cirque opens outward. Here the slope of the ice is reversed and steepens toward the head of the tongue. The latter, which reaches down between the horns at an angle of startling steepness, may best be described as a succession of ice cascades. In summer it exhibits blue glacial ice except at the terminus, which is blackened with englacial débris.

The trough occupied by the tongue is so steep and narrow that it

affords no lodging for moraines. When visited in 1931, the terminus rested directly on the smooth bedrock floor of the trough, and at the east side the exposed bedrock extended several hundred feet beyond the ice before passing beneath any accumulation of morainal material. The stream from the glacier empties into Leigh Lake, and the northwest corner of the latter is characteristically blue-green from the glacial flour which it receives.

FIG. 8.—East face of Mount Moran and Skillet Glacier, seen from near town of Moran. Jackson Lake and the moraines which surround it appear in the foreground. (Photo by Crandall.)

The position and character of the terminus of this glacier are determined largely by the calving of ice masses from the snout, rather than by melting, as in the case of the other glaciers described in this paper; consequently, they vary notably from year to year. The writer would estimate that for the period 1926–31 the highest position of the terminal cliff has been at least 400 feet above the lowest position. Ice caves have been noted in some seasons, and totally absent in others.

SKILLET GLACIER

Skillet Glacier is the conspicuous ice field on the east side of Mount Moran (Fig. 8), facing Jackson Lake and familiar to all who have viewed this mountain from the vicinity of the village of Moran. It is

a close neighbor of Falling Ice Glacier, being separated from it only by the easterly of the two "horns" previously described. In the Pinedale stage the adjacent heads of these glaciers were continuous across the divide behind the horn, and the latter must have stood above the ice as a conspicuous "cleaver."

Like its neighbor, Skillet Glacier occupies a cirque high on the face of Mount Moran, and similarly no distinct moraines encircle its margin, though a long train of boulders extends down the gulch beneath.

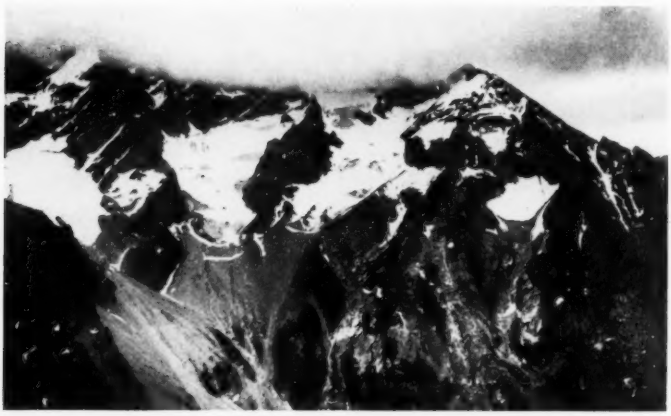

FIG. 9.—Triple Glaciers, photographed in 1930 from Bivouac Peak. The summit precipices of Mount Moran are beneath the clouds. (Photo by Theo. Koven.)

As compared with the cirque of Falling Ice Glacier, that of Skillet Glacier is broader and shallower. Probably because of the resultingly less effective protection given it, the glacier is of weaker development than the others of the range, consisting merely of an elliptical ice field without any terminal extension. From its head a steeply inclined, snow-filled couloir 1,500 feet or more in length extends upward and toward the right—the "handle" of the "skillet." Another persistent feature of the glacier is an exposure of bedrock which, in late summer, appears in the upper portion of the field, just below the point of attachment of the "handle."

At the time of its fullest extent in the Pinedale stage this glacier extended down to and beyond the base of Mount Moran, contributing

its quota to the broad mass of ice which, fed likewise by other glaciers issuing from canyons farther north, collected in the northern end of Jackson Hole and gave origin to the basin of Jackson Lake. The ancestor of Skillet Glacier was distinguished only in being the southernmost of the tributaries to this piedmont glacier.

TRIPLE GLACIERS

The large valley glacier which during Pinedale time occupied Moran Canyon is still represented by remnants, three ice fields which cling to the north face of Mount Moran at the foot of the great summit precipices (Fig. 9). Because of their close proximity and relationship, these are together designated the Triple Glaciers. They are not easy of access and as yet have not been explored; but they may be viewed from a distance, either from points northeast of Jackson Lake or, to much better advantage, from Mount Bivouac.

In Pinedale time these ice fields constituted the compound head of a glacier on Mount Moran which was tributary to the glacier of Moran Canyon. With the wane of glacial conditions, however, lower Moran Canyon became free of ice and the tributary on the north slopes of Mount Moran was isolated. Continued retreat led, eventually, to the dismemberment of its three heads to form the now distinct Triple Glaciers. Were conditions favorable to glacial growth to return, these sister-glaciers would soon become united; in fact, very slight readvance would suffice to bring about union of the middle and westerly glaciers.

Although the slopes beneath the three glaciers are exceedingly steep, cone-shaped morainal accumulations have formed beneath each. These are hundreds of feet high and, in their lower portions, overlapping. A short crescentic ridge, concave toward the ice, forms the apex of each. Below the moraines, discontinuous boulder trains extend down the lower slopes of the mountain almost to the floor of the canyon. Each of the Triple Glaciers has distinct lateral moraines, the west moraine of the westerly glacier being especially prominent. Study of these glaciers must be made at close hand before statements can be made relative to their more intimate features.[15]

[15] The writer expresses his obligation to Dr. C. K. Wentworth, who, in reviewing this report, offered helpful suggestions incorporated in the foregoing text.

COLLOIDAL PRIMARY COPPER ORES AT CORNWALL MINES, SOUTHEASTERN MISSOURI

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University of Chicago

ABSTRACT

Evidences are cited which indicate a hydrothermal ancestry for the mineralizing solutions. Some of the primary mineral species which suggest such an interpretation are: bornite, enargite, famatinite (?), isometric chalcocite, and primary covellite. Lack of alteration of the wall indicates very moderate temperatures. Several unusual types of structures and textures suggest ore deposition from colloidal solutions. A deep-seated, magmatic source is postulated for these solutions. The proximity of the mines to a major fault belt indicates that these faults may have facilitated the rise of mineralizing solutions from the postulated deep-seated source. The presence, a few miles to the southwest, of igneous intrusives may have significant bearing on the genesis of the ores.

INTRODUCTION

In an extensive literature devoted to the Mississippi Valley ore deposits widely divergent views have been advanced concerning the genesis of the ores. In studying the mineralization at the Cornwall mines the writer hoped to uncover sufficient diagnostic evidence to determine the genesis of the ores at these mines and thus fill a gap in the much larger picture of the Mississippi Valley deposits in general.

The mines are located about 10 miles southwest of the town of Ste Genevieve in southeastern Missouri (Ste Genevieve County) along a small stream tributary to the Rivier aux Vases (Fig. 1). Since 1872, nine years after copper was first discovered, mining has been carried on intermittently. The early history of the mines is given in a brief report by Nicholson.[1] In a more recent description of the workings Weller and St. Clair[2] reported the property as operating when last visited by them in 1916.

GENERAL GEOLOGY

The mines are situated on the northeast structural slope of the domelike Ozark uplift, the center of which lies about 30 miles to the

[1] Frank Nicholson, "Review of the Ste Genevieve Copper Deposits," *Trans. Amer. Inst. Min. Met. Eng.*, Vol. X (1882), p. 444.

[2] Stuart Weller and Stuart St. Clair, "Geology of Ste Genevieve County, Missouri," *Mo. Bur. of Geol. and Mines*, Vol. XXII (2d ser., 1928), p. 331.

southwest where pre-Cambrian granite outcrops over large areas. The rock formations involved in the mineralization at the Cornwall mines are the Cotter and Powell dolomites of Lower Ordovician age. For complete detailed description of the stratigraphy the reader is referred to Weller and St. Clair.[3] Just south of the mines is a structural belt characterized by a complex mosaic of normal faults which

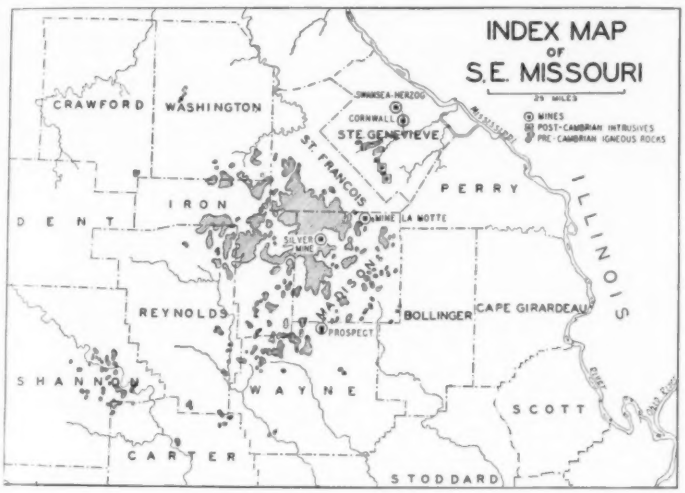


FIG. 1.—Map showing location of Cornwall mines

extends for 50 miles across the central part of Ste Genevieve County from northwest to southeast. Although the mineralization is not within the zone of strong faulting, some minor fractures of small displacement are observed in the mines. The main faulted zone extends southeastward and disappears beneath the flood-plain deposits of the Mississippi River. Its probable continuation may be seen across the river in the southern Illinois fluorspar district where the relations between faulting, igneous activity, and mineralization have been most recently studied by Bastin.[4]

A few miles to the southwest of the Cornwall copper mine, on the

[3] *Ibid.*, pp. 81–90.

[4] Edson S. Bastin, "The Fluorspar Deposits of Hardin and Pope Counties, Illinois," *Ill. Geol. Surv. Bull. 58* (1931).

opposite side of the fault zone, igneous intrusives cut the sedimentaries. Two of these have been studied in detail by Tarr and Keller,[5] and by Singewald and Milton.[6] As yet these have not been conclusively shown to be related to the mineralization of the region. Other intrusives in the locality are being studied.

ENVIRONMENT OF ORE DEPOSITS

GENERAL FEATURES

The ore occurs as irregular, discontinuous, lenticular to pockety blanket deposits, controlled by and developed along essentially horizontal stratigraphic breaks in the massive dolomite beds. The ore seems to be heaviest at the intersection of these horizontal levels, with minor vertical fractures along which some mineralization also has occurred. The vertical offshoots extend as much as 15 feet upward but only a few feet down from the horizontal ore levels.

Workings have been developed along three distinct horizons, and copper sulphides have been reported to occur along four additional levels, one outcropping and the other three revealed by drillings.[7]

While a number of horizons have been mineralized, extensive development has been principally along the stratigraphic break separating the upper and lower Powell members. This level is marked by thin alternating lenses of sandstone, conglomerate, and breccia.

SEQUENCE IN MINERALIZED ZONES

Development of older chert.—The chert of the Powell formation is not confined to the dolomite beds but occurs in considerable amounts as nodules in a sandstone member about 15 feet below the main mineralized level. This variety of chert has heretofore been termed "oolitic" due to the presence of variable amounts of included sand grains. It is being studied in more detail in an effort to confirm the writer's belief that it is syngenetic.

Dolomitization.—There is no apparent genetic relation between dolomitization and mineralization. The magnesium content of the

[5] W. A. Tarr and W. D. Keller, "A Post-Devonian Igneous Intrusion in Southeastern Missouri," *Jour. Geol.* Vol. XLI (1933), p. 815.

[6] Joseph T. Singewald, Jr., and Charles Milton, "An Alnoite Pipe, Its Contact Phenomena, and Ore Deposition near Avon, Missouri," *ibid.*, Vol. XXXVIII (1930), p. 54.

[7] H. Foster Bain and E. O. Ulrich, "The Copper Deposits of Missouri," *U.S. Geol. Surv. Bull. 267* (1905), p. 39.

Powell dolomite seems to be very uniform. Variation in crystallinity of the dolomite is not related to amount or locus of mineralization. Included fragments of dolomite in the heavily mineralized breccia are no more coarsely crystalline than elsewhere, as they should be if dolomitization were due to the mineralizing solutions.

Period of solution.—The breccia portions of the principal ore level have been interpreted by Weller and St. Clair[8] as being in part sedimentary in origin. The writer would recognize three breccia types, gradational with one another and all traceable in origin to extensive solution of the dolomite at the ore level. This solution, guided by, and advancing upward from, the original stratigraphic break, progressed unequally, resulting in large semicircular pockets up to 20 feet in diameter above the mineralized level. Subsequent collapse of a supporting column between adjacent solution pockets here and there has developed considerable crush breccia, some of which was modified by continued solution with the complete elimination of the dolomite and the further incorporation of the broken chert fragments in a fine, siliceous, insoluble residue. Silicification (chertification) of this matrix residue has resulted in a hard breccia composed entirely of chert (Fig. 2, *4*). This second-generation chert is apparently related to the mineralization. Nicholson[9] has illustrated and described the results of this solution in some detail.

PRIMARY MINERALIZATION

REPLACEMENT VERSUS FILLING OF OPENINGS

Compared to the total ore, the amount of sulphide ore directly and completely replacing dolomite is small. A large amount of dolomite has been removed and a smaller amount of sulphide deposited, but most of the sulphide does not show distinctive evidence of having been formed by replacement; instead, it seems to have filled open spaces already existent, such as spaces between sand grains and breccia fragments and vugs of highly porous dolomite. In those specimens which seem to show evidence of direct dolomite replacement by sulphides (Fig. 8, *5*), the percentage of dolomite within the sulphide area is nearly equal in many cases to the percentage be-

[8] *Op. cit.*, p. 333.

[9] *Op. cit.*, pp. 450–53.

yond the sulphide frontier (Fig. 8, *4*). Porosity in the dolomite is greatest near the sulphide area and decreases away from it. It is believed that this condition was caused by the incomplete solvent action of the mineralizing solutions advancing along grain boundaries. The sulphides, after coagulation and under the influence of surface

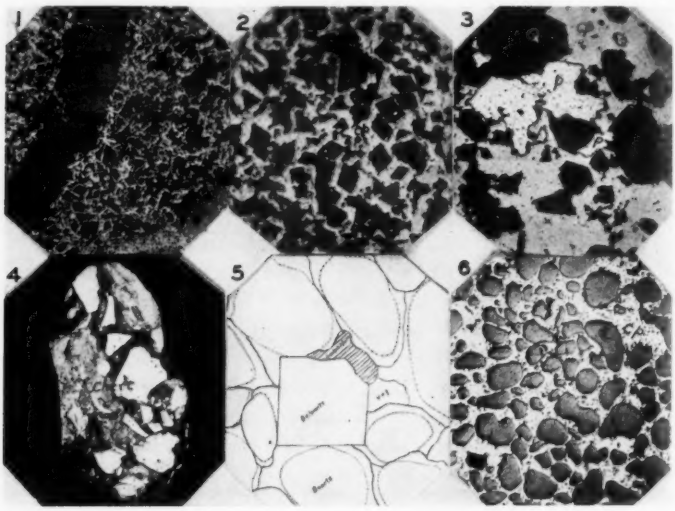


FIG. 2, ILLUSTRATIONS *1–6*.—*Cp*, chalcopyrite; *Ch*, chert; *P*, pyrite; *Q*, quartz. *1*. ×15. Chert pseudomorphic after dolomite. Note open shrinkage cracks developed in chert areas after the introduction of the chalcopyrite. *2*. ×56. Same as *1*. *3*. ×40. Quartz crystals developed by secondary addition rims on rounded sand grains. Note that quartz conforms to pyrite, which in turn conforms to original outline of rounded sand grain. *4*. ×$\frac{3}{4}$. Two generations of chert with breccia structure. The second generation (matrix) chert has developed by silicification of a very fine insoluble silica residue resulting from solution of the Powell dolomite. *5*. ×64. Dolomite crystals embedded in sandstone and automorphic toward secondary silica addition rims. *6*. ×15. Mineralized sandstone. Note separation of sand grains without replacement. In other less conservative fields the space between sand grains is as much as ten times the diameter of the largest sand grain. These mineralized areas are completely surrounded by unmineralized but silicified sandstone.

tension of the resulting gelatinous mass, were concentrated into a central zone in the total area affected by the solvent action of the disperse colloidal solutions.

COLLOIDAL VERSUS ELECTROLYTIC DEPOSITION

A wide variety of characteristic textures indicates clearly the colloidal condition of the ore material at the time of deposition. These textures are not limited to one or two species, but involve most of the primary minerals. Some of the textures and structures are widely

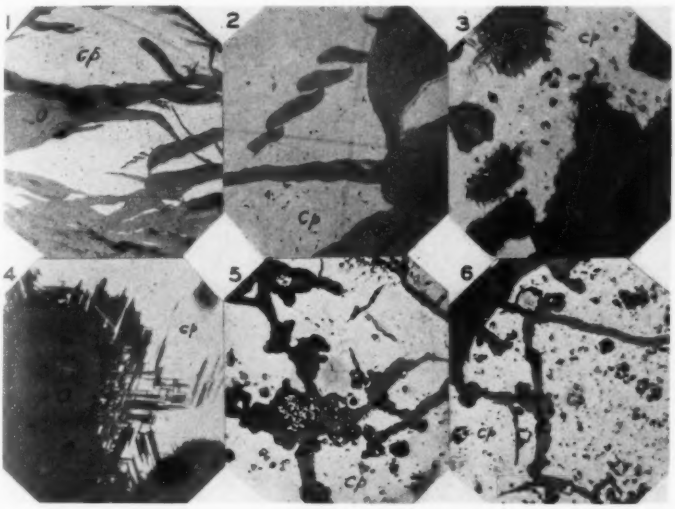


FIG. 3, ILLUSTRATIONS *1–6*.—*Cp*, chalcopyrite; *O*, intimate mixture of oxides of iron and copper; *Ma*, malachite; *Ca*, calcite; *P*, pyrite. *1*. ×20. Intimate mixture of copper and iron oxides replacing primary chalcopyrite. Note gross, clublike forms with smooth contacts as contrasted with characteristic ragged electrolytic replacement frontiers. These secondary oxides are believed to be colloidal in origin but not necessarily related to the primary colloidal sulphide deposition. *2*. ×100. Same as *1*. *3*. ×75. Replacement frontiers of secondary oxides controlled by crystallographic directions of host mineral—chalcopyrite. Malachite replaces the oxides. *4*. ×208. Same as *3*. *5*. and *6*. ×31. Calcite selectively replacing chalcopyrite and leaving pyrite. This replacement preceded enrichment.

accepted as characteristic of colloidally deposited material. Other no less well-defined textures, heretofore not recognized as colloidal in origin, have been so interpreted by the writer and are submitted for consideration as valuable additional criteria to be added to the scanty list of recognized colloidal textures now available. Some of

the most distinctive and most frequently occurring colloidally interpreted textures are described below.

1. Concentric, colloform structures. These are marked by (1) concentric shrinkage cracks in a single mineral (Fig. 4, *1*); (2) alternating bands of two different minerals as in Figure 4, *3;* (3) the

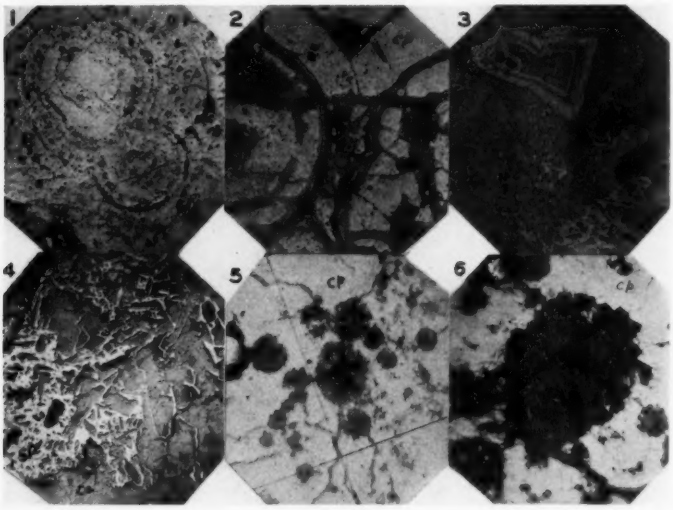


FIG. 4, ILLUSTRATIONS *1–6.*—*Cp*, chalcopyrite; *O*, intimate mixture of copper and iron oxides; *Ca*, calcite; *Sp*, sphalerite; *Cc*, chalcocite; *Co*, covellite. *1.* ×25. Colloform banding in primary chalcopyrite. Small white grains with high relief are pyrite; black areas are vugs. *2.* ×15. Colloform structures in primary ore controlling the development of secondary oxides. *3.* ×40. Alternating bands of calcite and sphalerite. These are believed to be essentially contemporaneous in age. *4.* ×32. Different field of same specimen as *3*. Under higher magnification many perfect crystals of calcite are seen to be entirely enclosed in sphalerite. *5.* ×200. Globules of an intimate mixture of primary isometric chalcocite and covellite showing cracked porcelain texture. *6.* ×400. Septaria shrinkage cracks developed in ex-solution(?) mixture of isometric chalcocite, and covellite. Same specimen as *5*. Note crystal faces of chalcopyrite against the globule.

controlling influence of inconspicuous primary structures on the development of secondary sulphides (Fig. 4, *2*); and (4) structures which, if not already visible, may be discovered by etching with acid fumes.

2. Syneresis shrinkage cracks. These may be classified into (1) "septaria" type of shrinkage expression involving small globular areas of one or two minerals (Fig. 4, *6*), or larger circular areas containing several minerals (Fig. 5, *3*, *4*, *5*, and *6*), (2) cracked porcelain textures mentioned by Lindgren[10] as being colloidal in origin (shown

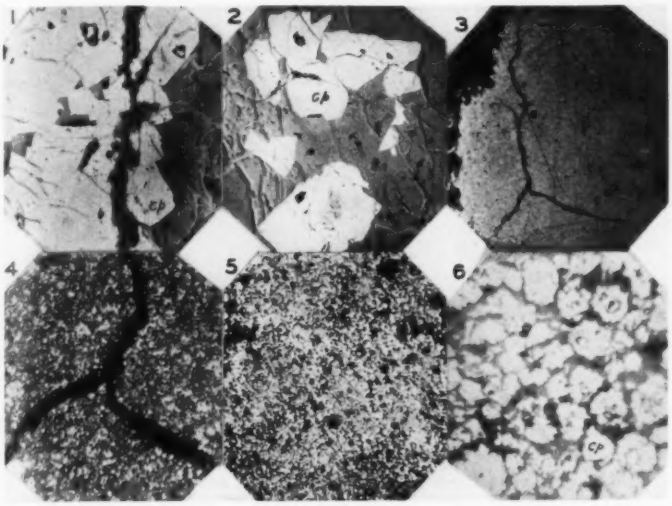


FIG. 5, ILLUSTRATIONS *1–6.—Cp*, chalcopyrite; *Bn*, bornite; *Cc*, chalcocite; *P*, pyrite. *1*. ×32. Primary bornite, interstitial toward automorphic chalcopyrite. *2*. ×30. Chalcopyrite crystals in primary bornite. Secondary chalcocite veinlets selectively replacing bornite. *3*. ×10. Septaria shrinkage cracks developed in a circular sulphide area (replacing dolomite pebble in sandstone). The cracks cut an intimate mixture of three primary minerals (see *4*, *5*, and *6*). *4*. ×40. Same as *3*, higher magnification. *5*. ×60. Less altered field of same specimen as seen in *3* and *4*. *6*. ×550. Same as *5*, high magnification. Note pyrite cores and bornite matrix to chalcopyrite spherules. Bornite partially replaced by secondary chalcocite (dark gray to black).

in small scale in Fig. 4, *5*); (3) lenticular syneresis cracks in definite angle "grid" pattern (Fig. 6, *5*). The cause of this pattern is not known, but it seems possible to the writer that it may be in part external and due to rotational stress applied to a semi-rigid con-

[10] Waldemar Lindgren, *Mineral Deposits* (4th ed.), p. 842.

tracting gel mass.[11] If the extensive solution at the mineralized horizon was accomplished by the mineralizing solutions, we can easily imagine unequal stress resulting from collapse. The lenticular character of the cracks is to be emphasized. (4) Lenticular cracks which

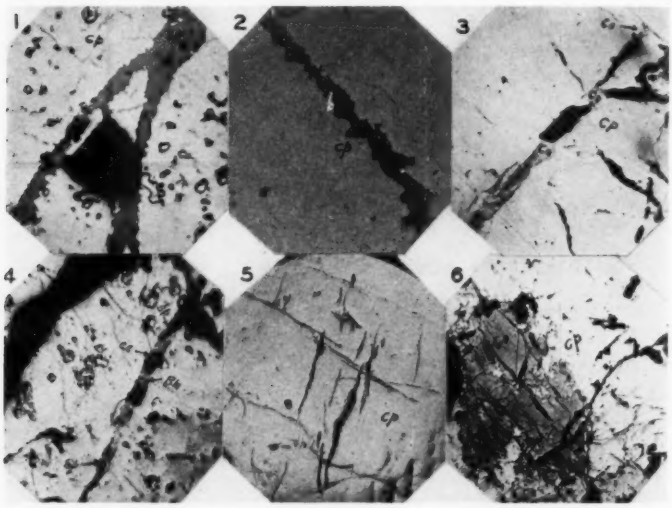


FIG. 6, ILLUSTRATIONS *1–6.*—*Cp*, chalcopyrite; *Sp*, sphelerite; *Co*, covellite; *Cc*, chalcocite; *Bn*, bornite. *1.* ×52. Sphalerite veinlets cutting chalcopyrite and showing crystal faces of host against guest. *2.* ×240. Same as *1*. This type of veinlet is believed to have been formed by injection of sphalerite along shrinkage or rupture fractures in semi-rigid chalcopyrite gel, with subsequent crystallization and development of crystal faces of chalcopyrite against the later sphalerite. *3.* ×40. Segmented veinlets of primary covellite alternating with open cracks (see *4*). *4.* ×40. Segmented veinlets. Alternating segments of sphalerite, bornite, chalcocite, and open vugs. This type of veinlet is believed to have been formed by expulsion of the disperse phase as chalcopyrite gel shrunk and formed cracks. Coagulation and contraction of the disperse material in the cracks caused segments of different minerals to alternate with each other and with vugs. This segmentation is an expression of selective coagulation of different disperse materials. *5.* ×10. Lenticular syneresis cracks with a definite angular orientation. *6.* ×50. Syneresis cracks controlled by the crystallographic directions in primary covellite.

have no definite orientation; and (5) lenticular cracks which may have a crystallographic control (Fig. 6, *6*).

[11] The writer is aware that sulphide coagulums are not true gels but better termed "gelatinous precipitates." Behavior of these precipitates after flocculation, however, follows closely the habits of gels. For simplicity, then, they will be spoken of as "gels."

3. "Framboidal"[12] clusters of tiny pyrite cubes and grains, the whole with a spheroidal outline. In some cases the grains have a matrix of primary chalcopyrite (Fig. 7, *4*), and in other cases the matrix is secondary (Fig. 7, *2* and *3*). In Figure 7, *1*, the spheroids

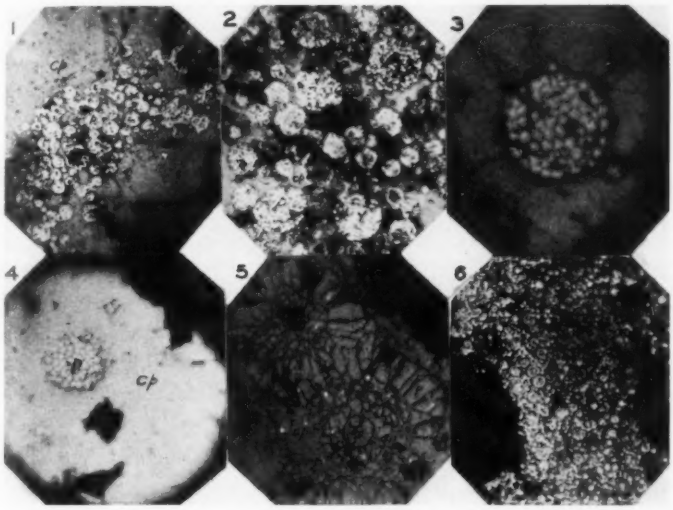


FIG. 7, ILLUSTRATIONS *1–6*.—*Cp*, chalcopyrite; *P*, pyrite. *1*. ×108. Solid spheroidal masses of pyrite in chalcopyrite. *2*. ×166. "Framboidal" spheroidal masses of pyrite in chalcopyrite due to metacolloidal crystallization. *3*. ×580. "Framboidal" clusters of pyrite grains due to crystallization of a gel globule. Secondary minerals separate the discrete crystals of pyrite. *4*. ×450. Same as *3* except that there has been no enrichment of the ore and hence chalcopyrite still forms the matrix of the pyrite cubes and grains. *5*. ×400. Radial "bomb" type of texture developed in pyrite with a small "framboidal" nest of pyrite grains in the center of one pyrite area. Chalcopyrite matrix of "bomb" fragments. *6*. ×64. Spheroidal globules of chalcopyrite, with pyrite core, and chalcocite (after bornite) matrix.

are composed of solid pyrite. There are present complete gradations between the solid spheroids and globules composed entirely of tiny uniform-sized pyrite cubes. These are probably metacolloidal in origin, having been formed by the crystallization of a globule of pyrite gel, crystallization starting at many points at the same time. In an alternative hypothesis, the texture might be considered to have

[12] After *framboise* (Fr.) for "raspberry."

been formed by the bunching of tiny pyrite grains as they floated in a chalcopyrite gel. On the basis of uniformity of grain size, gradational types between the grain clusters and solid spheroids, and the fact that in non-circular, irregular pyrite masses, the "framboidal" texture may be produced artificially by etching with nitric acid, the writer prefers the former interpretation.

4. "Frog-egg"[13] masses of chalcopyrite with pyrite cores, the whole implanted in a matrix of primary bornite and its alteration product chalcocite (Fig. 5, *5* and *6*, and Fig. 7, *6*).

5. Pyrite occasionally shows a radial habit with a nest of "framboidal" granules as a core (Fig. 7, *5*).

6. Veinlets of a late primary mineral cutting an earlier primary mineral with the walls of the veinlet composed of automorphic crystal faces of the host projecting into the guest (Fig. 6, *1* and *2*). These veinlets are believed to have formed by fracturing of a semi-rigid gel, either by shrinkage or by distortion, followed by injection of later colloidal material and subsequent crystallization of both with the host mineral assuming euhedral relations against the vein material.[14]

7. Segmented veinlets. The above-described veinlets, consisting dominantly of sphalerite, are interrupted in places by segments of bornite, chalcocite, and by open gaps, alternating along the course of the veinlets (Fig. 6, *4*). In another specimen lenticular shrinkage cracks are partially filled with covellite, constituting abrupt covellite segments alternating with gaps (Fig. 6, *3*). The coagulation of a disperse colloid to form a gel is followed by contraction (syneresis) with the expulsion of the disperse medium.[15] In the case of selective coagulation, the expelled disperse phase might be expected to carry

[13] Term suggested as descriptive of this peculiar texture as shown in Fig. 5, *5* and *6*. Note chalcopyrite spheroids carry a pyrite kernel and are enveloped in a bornite medium, an arrangement not unlike clusters of frog eggs.

[14] In a conversation between the writer and Dr. E. S. Bastin the latter agreed with the writer's interpretation and pointed out the similarity between these veinlets and some noted by him in ores from Cobalt, Ontario, and expressed the belief that the Cobalt occurrence should be explained in the same way. This agrees with the opinion of Lindgren in his paper on "Metasomatism," *Geol. Soc. Amer. Bull. 36* (1925), pp. 247–62, in which he expressed the belief that the Cobalt ores were colloidally deposited.

[15] H. R. Kruyt and H. S. van Klooster, *Colloids* (1930), p. 250.

considerable disperse colloidal matter which would be confined to the shrinkage crack. When this vein material coagulated, it would likewise contract and only locally fill the original shrinkage crack. If more than one mineral had been left in the original expelled disperse medium, the minerals would alternate with each other and with gaps formed by shrinkage of the vein material. If we could see these segmented veinlets sectioned in the plane of the original shrinkage crack, we would doubtless note circular globules of the different minerals separated by considerable vug space. This arrangement tells us nothing of the relative ages (order of coagulation) of the different minerals in the cracks.

8. In the case of the mineralized sandstones and also the chert nodules formed in sandstone, the sand grains are notably separated (Fig. 2, *6*). There has been no replacement of the sand grains, and yet in extreme cases the grains are separated by an amount equal to several times their diameter. The cause of this separation is not entirely clear. It seems probable that the explanation is to be sought in the gel nature of the matrix material. Whether it be expansion due to crystallization, expulsion of sand grains during contraction of the gel, or selective replacement of an earlier matrix of the sand, is not clear. Beyond the sulphide frontier, the sand grains are in close contact. The writer is inclined to think it is a colloidal phenomenon but would hesitate at the present time to offer it as a criterion for such origin.

9. Under descriptions of individual mineral species, evidence will be presented bearing on the relative order of coagulation of those species. This order is believed to be an expression of the disperse colloidal condition at the time of introduction (see general conclusions).

10. Replacement frontiers of sulphides invading dolomites are characteristically bulbous, mamillary, or nodular in appearance (Fig. 8, *5*). This feature is to be contrasted with typical ragged, feathery replacement frontiers developed by electrolytic replacement. The explanation probably lies in the effect of the high surface tension of gels causing rounded smooth contacts with the invaded rock.

THE PRIMARY ORE MINERALS

The primary ore of the Cornwall mines and vicinity consists dominantly of the sulphides of copper, iron, and zinc, named in order of relative abundance. Of the primary mineral species identified, chalcopyrite is by far the most abundant. Primary minerals occurring

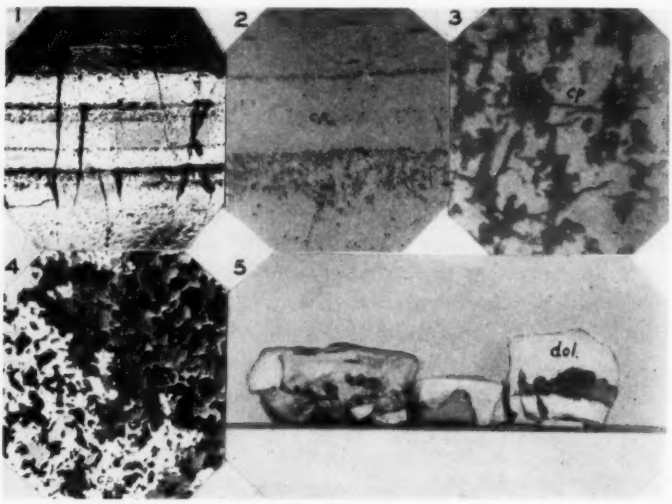


FIG. 8, ILLUSTRATIONS *1–5*.—*Cp*, chalcopyrite; *O*, oxides of copper and iron; *Dol*, dolomite; *S*, primary sulphides. *1*. ×12. "Mud-crack" type of shrinkage fractures developed in successive bands of chalcopyrite which was deposited on the wall of a fracture in dolomite rock (see *2* and *3*). *2*. ×40. Higher magnification of bands seen in *1*. See *3* and section in text on covellite for detailed explanation. *3*. ×292. High magnification of mottled bands seen in *1* and *2*. These darker oxide areas are dotted with replacement remnants of covellite and famatinite. This is believed to be essentially a primary texture. *4*. ×12. Characteristic porous dolomite beyond sulphide frontiers. Porosity developed by the action of the disperse solutions, but vacated by the contraction of the coagulated gel equivalent. *5*. ×$\frac{2}{3}$. Bulbous, gel type of replacement frontier of primary chalcopyrite invading dolomite. Note gray tinge to the sulphide areas due to high dolomite content as seen in *4*.

in lesser but important amounts, named in order of relative abundance, are: quartz, pyrite, sphalerite, marcasite, bornite, calcite, and dolomite. Rare occurrences of primary enargite, famatinite (?), chalcocite, covellite, galena, and fluorite were also noted.

Pyrite.—This mineral is rather uniformly scattered throughout the heavier ore as well as the slightly mineralized wall rock. Various types of occurrence include mosaic, coarsely crystalline intergrowth with marcasite, individual isolated euhedral cubes of variable size, individual grains with irregular outlines, solid masses with spheroidal

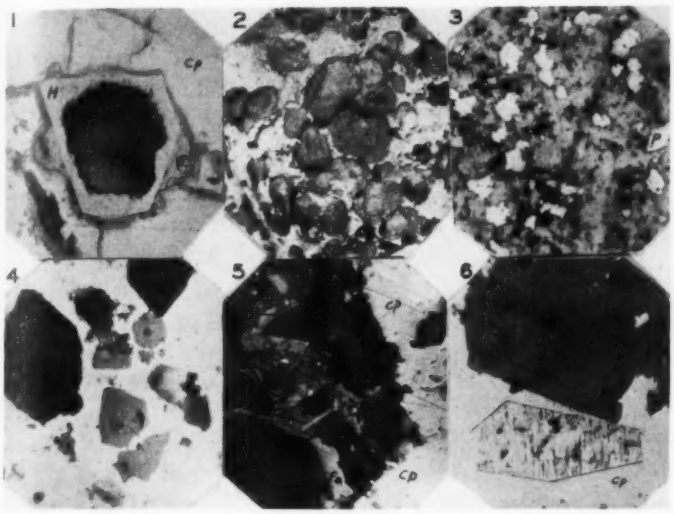


FIG. 9, ILLUSTRATIONS *1–6.*—*Cp*, chalcopyrite; *H*, hematite; *G*, goethite; *P*, pyrite; *Sp*, sphalerite; *Q*, quartz; *Co*, covellite; *Fa*, famatinite. *1.* ×125. Secondary hematite, pseudomorphic after pyrite and coated with drusy crystals of goethite. *2.* ×2½. Botryoidal chalcopyrite surfaces. *3.* ×160. Irregular pyrite grains scattered through chalcopyrite. *4.* ×150. Automorphic sphalerite crystals in chalcopyrite. This is in contrast to most sphalerite, which is later and conforms to chalcopyrite. *5.* ×140. Intergrowth of crystals of covellite and famatinite (?). *6.* ×100. Skeleton crystals of famatinite (?) in chalcopyrite.

outlines, "framboidal" clusters with spheroidal outlines, and radial spheroidal masses with or without "framboidal" cores. Perfect examples of the "framboidal" clusters may be developed by etching with nitric acid an irregular, apparently solid pyrite area, such as shown in Figure 9, *3*. Etching of the automorphic pyrite does not produce the granular texture. Gradational types between the solid spheroidal masses, the radial spheroidal masses, and typical "fram-

boidal" masses are quite common, and all these types are thought to represent ore which has crystallized from a gel condition. The automorphic pyrite is probably earlier than the chalcopyrite in which it occurs and probably represents preliminary sulphide deposition directly from the disperse phase before the proper electrolytic concentration had become sufficiently high to cause rapid flocculation and resulting gel textures. Pyrite is notably lacking in all sphalerite and calcite areas. See the section on colloidal textures and the illustrations for further information on pyrite occurrences.

Marcasite.—Types of marcasite occurrence include mosaic; crystalline; pre-gel intergrowth with pyrite and automorphic toward chalcopyrite; radial spheroidal masses implanted in chalcopyrite; and large radial, reniform masses in soft cherts and breccias and not closely associated with other sulphides.

Chalcopyrite.—This is by far the most abundant primary mineral. In the more massive type of occurrence it is dotted with pyrite and quartz grains. Where closely associated with other sulphides the textural relationships may be extremely intimate and characterized by spheroidal globules of chalcopyrite, many of which have "framboidal" clusters or irregular pyrite grains as a kernel. The circular globules of chalcopyrite are implanted in a groundmass of primary bornite. This arrangement would seem to suggest the order of coagulation or at least of crystallization as being pyrite, chalcopyrite, and bornite. A much coarser texture involving these three minerals in the same age relationship is seen in Figure 5, *1* and *2*. Here again pyrite is automorphic (not well shown in the pictures) and bornite is quasi-interstitial toward chalcopyrite.

Quartz.—Quartz is seen abundantly scattered through the ore as large hexagonal crystals with rounded sand-grain centers, as cryptocrystalline quartz rhombs, pseudomorphic after dolomite, and as finely crystalline quartz matrix (second generation chert) of an older chert breccia (Fig. 2, *1*, *2*, *3*, and *4*).

Sphalerite.—Very small amounts of sphalerite were noted as automorphic crystals implanted in massive chalcopyrite and as tiny drusy crystals lining vugs and cracks in the breccia (Fig. 9, *4*). Elsewhere massive sphalerite is seen as impregnation of porous dolomite and breccia and as veinlets cutting chalcopyrite as shown in Figure 6, *1* and *2*, and discussed under the section on colloidal textures.

Bornite.—This mineral occurs in small amounts but its presence is significant. It shows definite primary relationships with chalcopyrite (Fig. 5, *1*, *2*, *5*, and *6*). For further descriptions of its occurrence see the section on colloidal textures and the paragraph on chalcopyrite.

Calcite.—Normally calcite is seen replacing dolomite rock and lining cavities as drusy crystals. In an abnormal occurrence it has formed replacement veinlets in chalcopyrite. (Fig. 3, *5* and *6*). There is present a progressive sequence of crystal forms in the drusy calcite. At one end of the series, acute scalenohedrons are dominant. Later this type gradually gives way to rhombohedral forms with an acute rhombohedron as the final end member. Since this last acute rhombohedron was preceded by oxidation, it was probably deposited from descending solutions. According to Kalk,[16] this gradual change in crystal forms is to be expected as an accompaniment of temperature drop, the acute scalenohedral form representing low hydrothermal conditions.

Enargite.—This mineral is found only in very minor amounts, but has been identified in two closely related types of occurrence—as crystals lining vugs and as small isolated crystals enclosed in calcite areas. Identification by microchemical tests was checked by an X-ray spectroscopic analysis.[17]

Chalcocite.—In addition to the usual secondary orthorhombic chalcocite, there occur small amounts of primary isometric chalcocite in association with covellite, bornite, and sphalerite along segmented veinlets described under colloidal textures. In an equally characteristic occurrence it is noted as small spheroidal masses dotted with covellite blades and cut by septaria and cracked porcelain textures. Contacts between chalcocite and chalcopyrite are composed in large part of crystal faces of chalcopyrite (Fig. 4, *6*). Textural relationships between isometric chalcocite and covellite suggest ex-solution.[18] Experimental work by Pŏsnjak, Allen, and

[16] Georg Von Kalk, "Die Kristalltracht des Kalkspates in Minerogenetischer Betrachtung," *Centralblatt f. Min. Geol. u. Pal.*, Part A (1928), pp. 337–39.

[17] See acknowledgments.

[18] Alan Bateman and Samuel G. Lasky," Covellite–Chalcocite Solid Solution and Ex-solution," *Econ. Geol.*, Vol. XXVII, pp. 52–86.

Merwin[19] seems to have established the fact that chalcocite forming above 91° C. assumes the isometric form and that the presence of 8 per cent or more of dissolved covellite will prevent inversion to the orthorhombic form. This low temperature agrees with the evidence for hydrothermal origin of the solutions as well as the lack of alteration of the wall rock.

Covellite.—Primary covellite is minor in amount, but does occur in the ex-solution type of texture mentioned above as isolated grains in sharp contact with chalcopyrite, in segmented veinlets in chalcopyrite (not replacement veinlets), and as small automorphic grains intergrown with famatinite (?) (see illustrations).

Perhaps the most unusual occurrence of covellite, and the most unusual textural relationship in a deposit replete with extraordinary textures, is what may be termed a colloidal "graphic-like" intergrowth of covellite (now largely replaced by oxides) and chalcopyrite. The intimate intergrowth areas occur as bands which transect the specimen parallel to the surface of the chalcopyrite, which itself is cut by a "mud-crack" type of shrinkage expression (Fig. 8, *1*, *2*, and *3*). These mottled intergrowth bands are believed to represent periods during ore deposition when the carbonate content of the solutions became high enough to cause flocculation of the less easily coagulated covellite[20] along with chalcopyrite. Owing to fluctuation of introduction of mineralizing solutions, pure chalcopyrite was flocculated at certain times, and covellite and chalcopyrite were flocculated simultaneously at other intervening times. This explains why the mottled bands come on gradually and end abruptly when the delicate electrolytic balance was upset by a fresh surge of mineralizing solutions. At such a time chalcopyrite deposition continued or was soon resumed.

Famatinite.—This mineral, occurring in small amounts, has not been conclusively identified. It may be the pink variety of enargite known as luzonite. In view of its age difference and mode of occur-

[19] E. Pŏsnjak, E. T. Allen, and H. E. Merwin, "The Sulphides of Copper," *ibid.*, Vol. X (1915), pp. 491–535.

[20] C. F. Tolman and J. D. Clark, "Copper Sulphides from Colloidal Suspensions," *ibid.*, Vol. IX (1914), p. 576; J. D. Clark and P. H. Menaul, "Rôle of Colloidal Migration in Ore Deposits," *ibid.*, Vol. XI (1916), pp. 37–41.

rence, however, it is believed to be famatinite. While it may be seen in almost any specimen, it could not be isolated in sufficient quantities to make a positive identification. In its most characteristic occurrence it is observed as skeleton crystals in chalcopyrite (Fig. 9, *6*). It also occurs intimately intergrown with crystals of what is believed to be primary covellite (Fig. 9, *5*).

MODIFICATIONS BY SECONDARY ENRICHMENT

While secondary minerals occur in considerable amounts, little or no enrichment has occurred. This fact is probably largely due to the attitude of the ore bodies—dominantly horizontal with minor vertical offshoots of small dimensions.

Secondary minerals that have been identified include chalcocite, covellite, hematite, goethite, cuprite, malachite, smithsonite, cerussite and gypsum. At the upper mineralized level secondary alteration has followed the more usual habits of electrolytic solutions affecting copper-iron sulphides. At the lower horizon, however, the replacement frontiers are sharply contrasted with normal electrolytic replacement contacts (Fig. 3, *1* and *2*). These gross, clublike forms with rounded, smooth contacts are believed to be deposited from colloidal solutions. This is borne out by microchemical and microscopic studies, which seem to indicate intimate mixtures of copper and iron oxides for these secondary minerals replacing chalcopyrite directly.

CONCLUSIONS

ORIGIN OF SOLUTIONS

The deposit is interpreted as one of hydrothermal origin, so far removed from the parent igneous source that the temperature at the time and place of deposition had dropped nearly to that of ground waters. Mineral species, however, prove thermal ancestry. The unusually low temperatures account for the scarcity of high-temperature mineral species in the ore as well as the lack of influence of the solutions on the wall rock.

Positive evidence for the hydrothermal origin of the solutions is not overwhelming. A number of converging, though singly not conclusive, lines of evidence, however, seem to support this interpretation. Some of the facts which have been mentioned, and which sug-

gest an igneous ancestry, are: (1) primary bornite; (2) occurrences of enargite; (3) occurrences of probable famatinite; (4) good evidence of primary isometric chalcocite; (5) chalcocite-covellite solid solution and ex-solution textures; (6) primary covellite; and (7) sequence of calcite crystal forms.

ORE DEPOSITION

The mineral succession tables listed below are based on the more usual criteria for the determination of age relations. The writer wishes to point out, however, that this is not a usual type of mineral introduction, and age relations and criteria for the determination of age relations may not hold. In the opinion of the writer, the ores were introduced as dispersed and intimately diffused colloids of different mineral species and flocculation and a gel condition resulted. Order of flocculation of mineral species from such an intimate colloidal mixture would be a specific property of the mineral in question rather than an indication of the order of introduction.

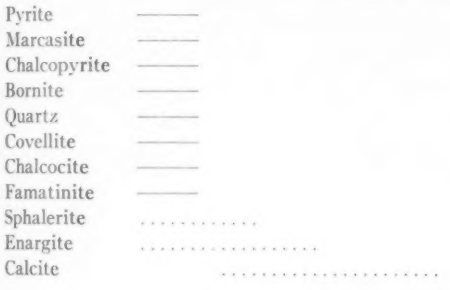
In the case of the injection veinlets of sphalerite cutting chalcopyrite (Fig. 6, *1* and *2*), there can be no doubt that the vein material was formed late, though it is not entirely clear whether this age difference is an expression of the late flocculation or late introduction. Crystallization of the host from a gel condition undoubtedly occurred after the injection of the guest. With the possible exception of calcite, sphalerite, and enargite, the writer believes that all of the materials now associated were introduced as a colloidal mixture at the same time, and that their present apparent age relationships are an expression of the order of ease of coagulation of species from that common mixture rather than an indication of the relative period of introduction. Experimental work by Tolman and Clark[21] and Clark and Menaul[22] suggested that certain sulphides coagulate from a dispersed colloidal condition in a definite order, that order being the reverse of ease of dispersion. Several of the copper sulphides studied by them occur in the Cornwall ores and in definite age relations. The order of age of these is pyrite, chalcopyrite, bornite, supposed famatinite, covellite, chalcocite. It is significant that this is the order of coagulation of these minerals as determined by the above-mentioned

[21] *Op. cit.* [22] *Op. cit.*

experimental workers, with the exception of famatinite, which takes the position of enargite. At the Cornwall mines enargite is intergrown with calcite and is quite late in development. It does not, however, show any evidence of having been deposited from colloidal condition. Its late age may be due to the fact that it was deposited from a true solution along with calcite. Primary galena occurs but in such small amounts that its age relations with respect to the dominant sulphides could not be determined.

In interpreting the age relations of the primary mineral species, two mineral succession tables were used. This seemed necessary in view of the colloidal origin and the resulting gel condition which obtained. The first table gives the ages with respect to the time of introduction; the second is in terms of period of flocculation from colloidal condition.

ORDER OF INTRODUCTION



ORDER OF FLOCCULATION

* May have been deposited from true solutions.

These succession tables were prepared with very great care after many specimens had been studied. While the tables are significant and tend to throw light on the general problem, the conclusions drawn therefrom must be made, as stated above, with recognition of the colloidal condition at the time of introduction. An interesting speculation arises when an attempt is made to offer an explanation of the observed order of flocculation of different mineral species from the disperse condition. Cox, Dean, and Gottschalk[23] believe the order is in the main due to degree of dispersion, but is influenced by many other factors. The order of flocculation determined experimentally by Tolman and Clark and Clark and Menaul was verified by Lasky[24] at Kennecott in the Copper River district of Alaska, and by the writer at the Cornwall mines. If degree of dispersion determines the order of flocculation, it is difficult to understand the coincidence of age relations between two naturally occurring, unrelated mineral suites, and between those and laboratory determinations on the same suite of minerals. It also seems more than mere coincidence that the order should be that of decreasing iron and increasing copper content in the mineral compositions. It seems probable that the order of flocculation is closely tied up with the charge on the disperse particles and the relation between that charge and the copper-to-iron ratio in those particles.

Summary of evidence for colloidal deposition.—1. Concentric, colloform structures.

2. "Syneresis" shrinkage cracks, septaria structures, cracked porcelain textures, etc.

3. "Framboidal" clusters and similar related pyrite habits and textures, such as radial pyrite.

4. Globular areas of chalcopyrite with pyrite cores, the whole occurring in a groundmass of bornite ("frog-egg" texture).

5. Veinlets with crystalline borders.

6. Segmented veinlets.

[23] Cox, Dean, and Gottschalk, "Studies on the Origin of the Missouri Cherts and Zinc Ores," *Univ. of Mo. School of Mines and Met., Tech. Series*, Vol. III, No. 2 (1916), p. 25.

[24] S. G. Lasky, "A Colloidal Origin of Some of the Kennecott Ore Minerals," *Econ. Geol.*, Vol. XXV (1930), p. 737.

7. Separation and pushing apart of sand grains.

8. Coincidence of order of coagulation and order determined experimentally.

9. Nodular, bulbous, or mamillary replacement frontiers.

10. Banded, colloidal "graphic-like textures."

CHARACTER OF SOLUTIONS

Some suggestion as to the probable nature of the mineralizing solutions is afforded by the consideration of certain facts. The amount of solution of the wall rock (dolomite) connected with the mineralization suggests the possibility that the solutions may have been slightly acidic. The original dispersed condition of the sulphides was probably due to and maintained by the presence of an excess of hydrogen sulphide.[25] In certain cases owing to hydrolysis, hydrogen sulphide solutions are alkaline in reaction. In regard to carbonate, however, hydrogen sulphide in solution acts like an acid. The acid condition of the solutions due to an excess of hydrogen sulphide may reasonably have been a retention of conditions existing when the solutions left the magmatic source. Ore-bearing, hydrogen-sulphide–charged emanations from magmatic sources are not uncommon. Clark and Menaul also showed that mineral colloids dispersed by H_2S were precipitated by a piece of calcite when that mineral was suspended in the solutions. The writer noted at the Cornwall mines, in a few specimens of chert breccia carrying an occasional fragment of dolomite, that there is a very definite suggestion of the movement of sulphides out of the breccia migration channel and a precipitation on the dolomite with some replacement. The writer believes that dolomite was the flocculating agent, and the conditions may have been very similar to those created in the laboratory by Clark and Menaul. The mineralizing solutions carrying the disperse colloids probably rose relatively rapidly along vertical fractures and spread more slowly along the sandstone-conglomerate stratigraphic break. Extensive solution of the dolomite created a sufficient concentration of the electrolyte to cause flocculation of the dispersed sulphides.

[25] Clark and Menaul, *op. cit.*, pp. 37–41.

Boydell[26] calls upon a probable excess of carbon dioxide as an effective agent for putting the carbonate electrolyte into solution and thus bringing about coagulation. He also postulates dispersion by the peptizing influence of excess hydrogen sulphide. The effect of hydrogen sulphide on the carbonate is not considered. An aqueous solution of hydrogen sulphide compares favorably with carbon dioxide in solution as a solvent for carbonates. Soluble calcium hydrosulphide and calcium bicarbonate are formed. The reaction proceeds to the right according to the equation $2CaCO_3+2H_2S \rightarrow Ca(HS)_2+Ca(HCO_3)_2$. Both compounds formed are fairly soluble and are effective coagulating agents. The writer sees no necessity for postulating the presence of carbon dioxide in the solutions, since hydrogen sulphide is already required as a peptizer in order to complete the process. By way of verifying the validity of the foregoing statements the writer polished a specimen of pure calcite, placed it in a bottle of distilled water, and bubbled H_2S through it for about five minutes. The bottle was then corked and allowed to stand for four hours. At the end of that time the polished surface of the calcite was examined and found to be deeply etched. Barium chloride was added to the solution in order to determine whether the etching was due to the formation of sulphuric or sulphurous acids. No sulphate or sulphite ions were detected.

If the foregoing reaction is acceptable as an explanation of the carbonate solution at the Cornwall mines, we have two simultaneously operating causes of coagulation of dispersed sulphides. One is the removal of the protecting agent, hydrogen sulphide, by a chemical reaction, and the other is the consequent production of two effective electrolytes as a result of the same chemical reaction.

Since thermal ancestry and low-temperature conditions are postulated, a distant magmatic source is necessarily inferred. How hydrothermal mineral-bearing solutions, passing up through a zone with a wide range of temperature-pressure conditions, could escape mineral deposition constitutes a real question. The explanation is probably to be sought in a consideration of the dispersed colloidal condition of the mineral matter at the time of introduction. If we postulate the

[26] H. C. Boydell, "Operative Causes in Ore Deposition," *Trans. Inst. of Min. and Met.*, Vol. XXXVI (1927).

material being kept in disperse condition because of the presence of an excess of hydrogen sulphide, we can call upon five means, according to Boydell,[27] of inducing relatively rapid deposition of sulphide material. These coagulating causes he lists as (1) influence of electrolytes; (2) mutual reaction of two disperse phases; (3) removal of protecting agent; (4) change of temperature or/and pressure; and (5) electrolysis.

Of these the writer need appeal to but one, and recognize the probability of the partial effect of two others. One of the latter two is the expected decrease of temperature and pressure consequent upon the rise of the solutions toward the surface. The other is the loss of hydrogen sulphide. This could presumably be accomplished either by mingling of the mineralizing solutions with ground water and a resulting dilution of the hydrogen sulphide content to cause coagulation, by a gradual diffusion of hydrogen sulphide under the influence of lesser pressure conditions or through removal by reaction with the electrolyte, as discussed above. A third alternative, and perhaps the most effective, would be the coagulation of the disperse material when the solutions came into contact with an effective soluble electrolyte. The carbonates of the Powell-Cotter dolomites, as well as the small amount of aluminous material along the slight unconformities, would serve as adequate coagulators. As already stated in discussing the mineralized portions of the chert-dolomite breccia, there is definite evidence of a selective coagulation of sulphides on the dolomite fragments.

It seems plausible to the writer to think of these sulphides of colloidal origin as differentiates from a system which at its magmatic source was rich in truly dissolved sulphides and lean in dispersed colloidal sulphide material. While in transit through probable igneous wall rock there would be precipitation of mineral matter from true solutions due to temperature-pressure changes and a resulting relative concentration of the colloidal phase in the solutions, since the peptizing influence of hydrogen sulphide would prevent deposition of the finely suspended material. If the original solutions at their source were only moderately rich in hydrogen sulphide, we might conceive of certain difficultly dispersible sulphides being deposited

[27] *Ibid.*, p. 42.

near the source, while those species easily dispersed would move away and perhaps be completely segregated. However, there might well be slight overlaps, and small amounts of the colloidal phase might be included in the minerals deposited from true solutions. And again, if an easily coagulated material is selectively deposited, it might be expected to carry with it some of the disperse material as minute colloidal inclusions. There would, then, be complete gradations from (1) deposits of dominantly colloidal origin, with a negligible amount of material deposited from true solutions to (2) deposits dominantly derived by crystallization from high-temperature solutions, with a small amount of colloidal material hidden within growing crystals. We may in consequence have certain species of minerals which form by crystallization from true solutions closely associated with other species which exhibit every evidence of colloidal deposition. Again, we may expect some species to form into automorphic crystals directly from the disperse colloidal condition as an expression of specific ability, or we may have minerals which possess ability to go from colloidal gel condition to automorphic crystalline condition without retaining much evidence of the original gel state. Several of these specific variations are suggested in the Cornwall copper ores. Pyrite seems to have an ability to form crystals from the disperse colloidal condition as well as to crystallize from the gel condition. Etching of chalcopyrite frequently brings out crystalline texture with individual crystals transecting banded colloidal structures. Relict structures demonstrating colloidal ancestry are not always present. The fact that any are preserved is probably due to the unusually mild conditions of deposition. Under more usual moderate or extreme temperature-pressure conditions metacolloidal crystallization would be so perfect and complete that it would be exceptional to find colloidal textures preserved even though it may have been the principal mode of ore deposition.

According to Van Bemmelen, von Weimarn, and others, there is a different precipitation concentration for each electrolyte and sol.[28] This concentration is very much greater when the electrolyte is

[28] J. M. van Bemmelen, "Die Absorption," *Gesammelte Abhandlungen ueber Kolloide und Absorption*, (Theodor Steinkoff, 1926); P. P. von Weimarn, in *The Chemistry of Colloids*, by W. W. Taylor (Edinburgh, 1915).

added slowly; also flocculation is less sudden. Before the rapid flocculation stage is reached there is nothing to prevent crystallization keeping pace with flocculation, particularly in the case of such minerals as pyrite, which are known to possess strong crystallizing properties. This in substance amounts to direct crystallization from the disperse phase. Von Weimarn also states that, whereas a definite concentration of electrolyte is necessary for rapid flocculation, a further increase in concentration of the electrolyte prevents rapid coagulation. Thus in cases where there is a slow, uniform increase in concentration of an electrolyte (as there would be if production of electrolyte were due to solution of a relatively slowly soluble carbonate wall rock), there is a well-defined concentration value for the rapid deposition of the disperse material. Preceding and following this stage, coagulation would be slow and crystallization might well keep pace with flocculation. This would create automorphic crystals of some of the most easily and least easily flocculated minerals. Thus is offered a possible explanation for the early pyrite and the late sphalerite being automorphic crystalline, while the bulk of the ore shows massive colloidal textures. This would also explain why the pyrite is always included in later chalcopyrite, while sphalerite is very commonly found as drusy vug linings or embedded in the healing-solution mineral calcite. Lindgren[29] says concerning the colloidal origin of the Cobalt ores:

> The complex gel was almost immediately transformed into crystalloids, resulting in the mixture of minerals which we now find. . . . . Because the crystallization followed so closely on the heels of gel-precipitation, it is not surprising that we find places where crystals were formed directly in the country rock or in the vein filling, but they do not appear to be the principal mode of ore formation.

Presumably conditions might be such that all of the disperse ore material might be deposited before the rapid precipitation concentration was realized, in which case we would have a colloidal ore deposit composed entirely of coarsely crystalline material with almost nothing to indicate its colloidal ancestry. Only during the rapid coagulation stage would identifiable colloidal textures and structures be developed, and even these might be eliminated by metacolloidal adjustment and crystallization under the influence of pres-

[29] W. Lindgren, "Metasomatism," *op. cit.*, p. 261.

sure and temperature. In the laboratory, Chrustchoff[30] produced quartz and tridymite crystals, both from silica sol and from silica gel, by heating in closed systems from 250° to 350° C. Tolman and Clark[31] formed chalcocite crystals from the disperse condition, and later Young and Moore[32] at a temperature of 30° C. readily produced crystals of chalcocite, covellite, chalcopyrite, and bornite by first dispersing with concentrated solutions of hydrogen sulphide.

The writer is very much in sympathy with the experimental evidence and theoretical considerations advanced by such workers as Bateman; Boydell; Clark and Menaul; Cox, Dean, and Gottschalk; Lasky; Lindgren; Tolman and Clark; and Schneiderhohn on the subject of colloidal ore deposits. The conclusions reached by Bateman and Lasky concerning the colloidal origin of the sulphide ores at Kennecott seem sound. Here they have described a type of mineral occurrence long suspected by a few workers, but previously not so conclusively linked with ascending solutions. The similarity between the Kennecott ores and those described above is very close. The mineralizing solutions apparently differed in but two respects. The iron and silica content of the Kennecott mineralizing solutions was very low, pyrite is conspicuously absent, and chalcocite is the dominant primary copper mineral; quartz is rare. At the Cornwall mines pyrite and marcasite are very abundant, and, as would be expected, chalcopyrite is the dominant copper mineral. So abundant and persistent was the iron content of the solutions that only small amounts of primary chalcocite and covellite are found.

There is no evidence of any reaction between an early coagulum and disperse material to produce a third mineral species, as described by Lasky.[33] He explains Liesegang diffusion bands of chalcopyrite and bornite as due to the diffusion of chalcocite molecules through the chalcopyrite coagulum with the resultant development of bands of bornite. Bornite associated with chalcopyrite at the Cornwall mines occurs as a matrix for globules of chalcopyrite or as interstitial mate-

[30] K. von Chrustchoff, *Neues Jahrbuch f. Min. Geol. u. Pal.*, Vol. I (1887), p. 205.

[31] *Op. cit.*, p. 573.

[32] S. W. Young and Neil Preston Moore, "Laboratory Studies on Secondary Sulphide Enrichment," *Econ. Geol.*, Vol. XI (1916), pp. 349–65.

[33] *Op. cit.*

rial between chalcopyrite crystals. There is every evidence that the bornite was coagulated from disperse condition as the bornite molecule. Its age relation is believed to be an expression of electronic charge on the disperse particles and is determined by the ratio of copper to iron in the molecule. It occupies its proper age position in a sequence of minerals of decreasing iron and increasing copper content.

During the progress of the study of the Missouri ores, the writer was impressed with the need of additional criteria for the recognition of colloidal textures. The conventional colloform structure, while present, was one of the least conspicuous. It is suspected by the writer that, owing to the lack of careful microscopic study, some deposits of colloidal origin may have been heretofore overlooked, and that certain unusual, uninterpreted, or ignored types of texture may prove to be colloidal in origin. The writer feels that the subject has not received enough attention, and believes further that zonal distribution of base metal sulphides may exist due to this colloidal mode of origin, as suggested by Tolman and Clark.[34] Sulphide deposits carrying essentially the same suite of materials, but differing widely in the ratio of these minerals to each other, may have a common origin and are but zonal expressions of the mineralization due to differential ease of dispersion and coagulation of the sulphides. Should one of a series of such deposits not show colloidal textures, we cannot assume a priori that all are not genetically related. As indicated above, ease of dispersion may just as effectively cause a species separation from a common suite as order of coagulation. One mineral may be deposited from true solution while the other more easily dispersed species may be carried on as a colloid under the influence of a peptizer such as H_2S. In this case the mineral first deposited might carry a few automorphic crystals of the second as an expression of slight crystallization from the disperse phase. The lead-zinc deposits of the Mississippi Valley, carrying some copper and associated regionally with deposits dominantly cupriferous, offer an example of a type of association which may eventually prove to be related to

[34] *Op. cit.*, pp. 37–41.

those here described. Emmons[35] has drawn an impressive picture of the zonal distribution of these ores. Cox, Dean, and Gottschalk have postulated a colloidal origin for some of the Tri-State ores, though they consider them as having been deposited from artesian waters. The Cornwall copper ores are definitely colloidal in origin and were deposited from ascending thermal solutions. It seems that any further attempt to bring the Mississippi Valley ore deposits into a single metallogenetic province should, of necessity, at least consider a colloidal explanation for the zonal distribution and segregation described by Emmons. We cannot assume that lack of dominance of colloidal textures and the existence of the idiomorphic crystalline expression of a suite of minerals conclusively indicate non-colloidal origin. It has long been known that crystallization directly from the disperse colloidal condition may take place. If it can be established that zonal distribution of sulphides exists as a function of the order of coagulation or dispersion of specific minerals, the significance and possible economic application becomes at once apparent.

ACKNOWLEDGMENTS

The writer is indebted to Professor Edson S. Bastin of the Department of Geology at the University of Chicago for many critical suggestions. Grateful acknowledgment is also made to Dr. C. W. Pierce of the Department of Chemistry at the University of Chicago for confirming the identification of enargite by making an X-ray spectroscopic analysis of the mineral.

[35] W. H. Emmons, "The Origin of the Deposits of Sulphide Ores of the Mississippi Valley," *Econ. Geol.*, Vol. XXIV (1929), pp. 221–71.

SUDBURITE, A METAMORPHIC ROCK NEAR SUDBURY, ONTARIO[1]

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ABSTRACT

"Sudburite" is the name applied to a rock composed essentially of plagioclase and pyroxene, with hornfels texture, occurring in proximity to the nickel intrusive near Sudbury, Ontario. Evidence is given to show that it is due to the recrystallization of greenstone and other rocks, older than the nickel intrusive, and that the instrusive was the cause of its formation.

INTRODUCTION

Certain rocks in the vicinity of the nickel intrusive, near Sudbury, Ontario, have been described by Coleman under a variety of names, as "micronorite,"[2] "older norite,"[3] and "sudburite."[4] In 1903 he described these rocks as a variety of the norite of the nickel intrusive, with reservations as to origin.[5] Later the sudburite was stated to be a basic extrusive. The present writer encountered this sudburite in the course of field work during 1928, and afterward, through microscopic examination, concluded it to be a metamorphic rock. The most abundant minerals in sudburite are monoclinic pyroxene, orthorhombic pyroxene, and plagioclase. The texture is fine-granular. The writer believes the important processes in the formation of this rock were the recrystallization of large plagioclase crystals into fine granules, and the crystallization of pyroxene from amphibole and biotite. Where these two processes have gone to completion, typical sudburite (plagioclase-augite-hornfels) is produced. In places plagioclase has recrystallized, whereas hornblende has not, which gives rise

[1] Published with the permission of the Director, Geological Survey of Canada, Department of Mines, Ottawa, but without concurrence in all the opinions expressed.

[2] A. P. Coleman, "The Northern Nickel Range," *Ont. Bur. of Mines 13th Ann. Rept.*, Part I (1904), p. 214.

[3] "The Sudbury Nickel Field," *Ont. Bur. of Mines 14th Ann. Rept.*, Part III (1905), p. 78.

[4] "The Proterozoic of the Canadian Shield and Its Problems," *Problems of American Geology* (1913), p. 103.

[5] *Ibid.*

to difficulty in naming this variety, since it has not all the typical features of sudburite. A further complexity is introduced in sudburite, because the pyroxene and frequently the other minerals also have been altered nearly everywhere to hornblende.

Sudburite occurrences are restricted to the vicinity of the outer margin of the intrusive. It was apparently formed from different types of rocks, but more strikingly from basic ones.

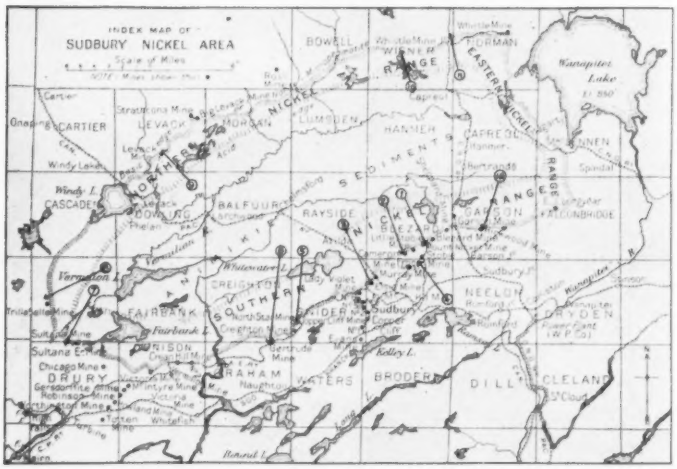

FIG. 1.—Map of the Sudbury Nickel-Copper area, Ontario, Canada, showing the location of the mines. The outlines of the norite-micropegmatite are also indicated.

OCCURRENCES AND DESCRIPTION

A sketch made of the Sudbury area shows the location of the places mentioned below (Fig. 1).

Locality 1, Blezard Township, lots 4 and 5, Concessions I and II.—On maps accompanying Coleman's reports, sudburite is shown as a somewhat elongated mass extending along the norite contact and up to about half a mile from it.

It is well exposed at the junction of the Sudbury-Capreol highway and the road leading off it to Mount Nickel mine, and again on the east side of the Sudbury-Capreol highway, south of the slag piles at Blezard mine. The fine-grained rock is traversed by raised amphi-

bole veinlets, and amphibole crystals (up to $\frac{1}{2}$ inch in diameter) with embayed outlines are found in it. Amphibolitic rocks, apparently formed by alteration of sudburite, intervene between it and the norite of the intrusive. The fine-grained sudburite grades into rocks in which the recrystallization processes, giving rise to sudburite, have not gone to completion, and also into rocks in which later alteration, largely to amphibole, has obscured its original nature. Thus, in addition to the typical fine- and even-grained variety there are others with abundant and, in many cases, large hornblende crystals. Accurate mapping of these rocks would require very detailed work.

Thin sections of fine-grained sudburite contain monoclinic pyroxene, orthorhombic pyroxene, plagioclase (between andesine and basic labradorite), and magnetite as the principal minerals. Biotite is present in small amounts. Quartz and pyrite are mentioned by Coleman[6] as minerals occurring in very small amounts. All these minerals are unaltered. In some thin sections pyroxene and plagioclase are in about equal amounts. Orthorhombic pyroxene is often more abundant than monoclinic.

A granoblastic texture is characteristic of the rock. The granules are usually of rather uniform size (Fig. 2), but in places pyroxene granules are larger than those of plagioclase. Occasionally pyroxene granules are aggregated to form bands across the thin section. Magnetite in places shows idiomorphic outlines against the other minerals. Thin sections of large amphibole crystals contained within, and having irregular outlines against, fine-grained sudburite show pyroxene granules developed in them. The amphibole crystals are remnants of the original rock not completely altered to sudburite. In places remnants of hornblende are surrounded by pyroxene granules (Fig. 3).

Locality 2, Blezard Township, lot 8, Concession IV.—Through the courtesy of Dr. T. L. Walker, the writer had the opportunity of examining some thin sections of rocks from this locality. Chemical analyses of these rocks have been given by Coleman, Moore, and Walker[7] in their important work on the nickel intrusive.

[6] "The Northern Nickel Range," *op. cit.*, p. 215.

[7] A. P. Coleman, E. S. Moore, and T. L. Walker, *The Sudbury Nickel Intrusive*, "Contributions to Canadian Mineralogy" (University of Toronto Press, 1929), pp. 28–29.

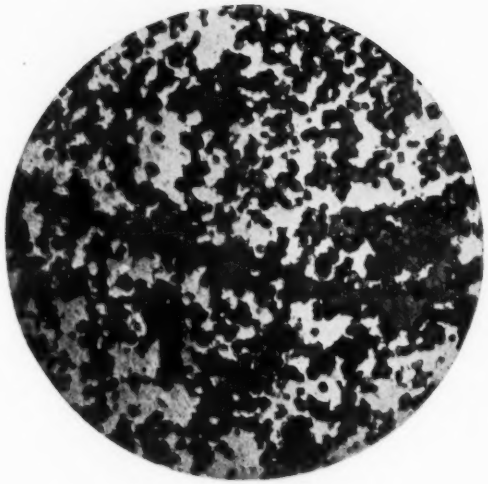

FIG. 2

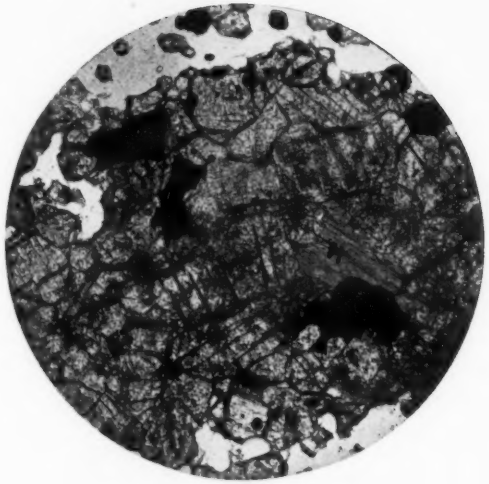

FIG. 3

FIG. 2.—Sudburite, near Blezard mine. Pyroxene and plagioclase granules. Amphibole veinlet, crossing section, alters pyroxene particularly. Ordinary light, ×7.

FIG. 3.—Sudburite. *H* = hornblende, surrounded by pyroxene granules, which appear to have formed from it. Ordinary light, ×25.

A glance at the analyses given shows that analysis No. 64 differs markedly from the rest of the series—Nos. 65, 66, etc. Under the microscope (Slide No. M–15956, Royal Ontario Museum, Toronto) the rock of this analysis proved to be sudburite.

Locality 3, McKim Township, lots 11 and 12, and adjacent areas of Snider Township, lot 1, Concessions IV and V.—This is in the vicinity of Murray mine. The writer saw typical sudburite here and also varieties with hornblende in varied amounts.

Locality 4, Frood Stobie offset.—The rock about 100 feet southeast of the shaft (formerly owned by the Mond Nickel Co.) on the offset and also that from immediately west of the open pit at Stobie mine were found, under the microscope, to be varieties of sudburite.

Locality 5, near Creighton mine.—A thin section (Slide No. M-15915, Royal Ontario Museum, Toronto) from here was identified as sudburite.

Locality 6, Creighton Township, lots 2, 3 and 4, Concession I (in the vicinity of Gertrude mine).—The sudburite here is in a quartz gabbro mass and is believed to be a metamorphosed part of it. Typical sudburite was not found in contact with norite (of the nickel intrusive) but separated by 20 yards or more. The intervening rock has been highly altered, largely to amphibole. About 300 feet west of the railway triangle in lot 3, Concession I, is a small area of sudburite.

Locality 7, near Sultana mine.—Parts of the greenstone at and near the mine are sudburite.

Locality 8, Trill Township, lot 9 (west side, south half), Concession III.—Inclusions in the granite here, as shown by thin sections, are similar to sudburite.

Locality 9, near Levack mine.—Brackenbury[8] reports the occurrence of sudburite in this locality.

Locality 10, Wisner Township.—Coleman[9] mentions its occurrence near Joe's Lake in this township.

Locality 11, near Nandair on the Canadian National Railway, north of the outer norite contact.—A thin section (Slide No. M-15922, Royal

[8] C. Brackenbury, "Notes on the Rocks at Levack," *Ont. Bur. of Mines 23d Ann Rept.*, Part I (1914).

[9] "The Sudbury Nickel Field," *loc. cit.*, p. 81.

Ontario Museum, Toronto) labeled "greenstone cement of granite fragments" is incompletely developed sudburite. A thin section of granite showed partial recrystallization of large albite crystals into smaller grains.

Locality 12, near Kirkwood mine.—The occurrence of sudburite here was reported by Coleman.[10]

Regarding the distribution of the older norite or sudburite, Coleman[11] states "it is not known to occur anywhere in the district at a distance from the nickel range, but always at or near the margin of the nickel-bearing norite." The present writer found no exception to this statement. Sudburite is distinguished from the norite of the nickel intrusive by the nearly complete absence of quartz, the small amount of biotite, the abundance of mafic minerals, and the granoblastic texture.

ORIGIN

That sudburite is merely metamorphosed basic rock is shown by its typical granoblastic texture, by the stages in the recrystallization of plagioclase to granules, at first in small amount and later completely, and by the occurrence of amphibole with pyroxene granules developed in it. Stages in the development of pyroxene in amphibole may be seen in which only remnants of the latter mineral are left, these being completely surrounded by the former. Pyroxene granules have also formed in biotite. In the early stages of this alteration the original texture of the rock is retained, but in typical sudburite it is lost or greatly obscured. The writer saw no suggestion of any process other than recrystallization in the breakdown of the plagioclase to granules, and the alteration probably occurred without important changes in the chemical composition of the rock as a whole. Schwartz[12] gives an account of alteration in greenstone in northeastern Minnesota at the contact with the Duluth gabbro, to a rock there called "muscovadite." This term was introduced for the granular altered greenstones, altered sediments, and gabbro phase along the gabbro contacts; but later work showed it to be a contact metamorphic rock of both greenstones and sediments. Through the courtesy

[10] *Ibid.*, p. 81. [11] *Ibid.*

[12] G. M. Schwartz, "The Contrast in the Effect of Granite and Gabbro Intrusion on the Ely Greenstone," *Jour. Geol.*, Vol. XXXII (1924), p. 89.

of Dr. J. T. Stark, of Northwestern University, the present writer had an opportunity of examining sections of "muscovadite" and found them almost identical with Sudbury types. Sudburite is believed to have an origin similar to that of "muscovadite."

Sudburite is a rock formed through high-grade metamorphism. The intense conditions leading to the formation of this rock appear to have arisen from the intrusion of the nickel intrusive. The general position of the sudburite—nowhere far from the outer contact—supports this interpretation, though it might be expected that sudburite would occur more constantly and regularly around the contact. More work is needed to substantiate the interpretation.

The Frood-Stobie offset is separated from the intrusive by granites younger than the norite, highly metamorphosed graywackes, and greenstones. The sudburite in the offset is about a mile from the present outer norite contact. This is farther than might be expected if its formation is caused by the intrusive. It seems plausible to assume that the offset and the intrusive were once much closer and were separated by the later granite intrusion.

Wandke and Hoffman,[13] in describing the alteration of basic inclusions in granite near Sudbury, give the breakdown of pyroxene and plagioclase to granules (giving essentially sudburite) as a stage in the granitization of the basic rock.

The writer did not examine rocks from the same place as those described by Wandke and Hoffman; but the sudburite, apparently similar to basic rock in a certain stage of alteration to granite, did not suggest a similar origin for it. Apart from the lack of any regular spacial arrangement of the granite, the type of change is different from that usually associated with granite intrusion. The only alteration of sudburite which the writer saw was to an amphibolite and not to a granite.

ALTERATION OF SUDBURITE

Alteration of sudburite by hornblende veinlets with preferential replacement of the pyroxene granules has been mentioned. In many places amphibole has replaced both pyroxene and plagioclase to a

[13] A. Wandke and R. Hoffman, "The Sudbury Nickel Deposits, *Econ. Geol.*, Vol. XIX (1924), pp. 189–190.

great extent. This amphibolitization is not confined to the sudburite; it has also gone on in the norite near and at its outer contact, and in granites as near Sultana mine.

RELATION TO MINERALIZATION

The time relation of the formation of sudburite to the nickel-copper mineralization is important. Three possibilities are presented: the recrystallization took place before, during, or after mineralization. It is quite possible that a relation valid for one place may not hold true for another. In the sudburite near Gertrude mine and in the Frood offset, small quartz stringers with amphibole edges carry pyrrhotite and chalcopyrite, and these are abundant in the rock adjoining the vein. The sulphides corrode and replace the silicates. Amphibole veinlets often occur in the sudburite without accompanying mineralization. The few observations of the writer suggested that the same agents which caused the formation of amphibole, also brought in the ore minerals. This indicates that mineralization is subsequent to recrystallization.

REMARKS

It appears plausible to regard the formation of sudburite and its later alteration to amphibolite as two effects of the nickel intrusive. The first effect presumaby arose from the change in physical conditions, principally the higher temperature, caused by the intrusive during and immediately after its intrusion in the rock adjacent to its outer contact. The second effect, after the intrusive had partly crystallized, was presumably caused by fluids emanating from it.

If the hypothesis which the writer advances for the formation of sudburite should prove correct, it may assist in solving some of the problems of the Sudbury area. Thus the apparently fresh lavas south of the Southern range are often puzzling to one unacquainted with the general geology of the Sudbury area, so much so that often one imagines an age difference between them and the more altered varieties (greenstones) close by. In reality the difference is due to a difference in metamorphism.

It appears to the writer that sudburite has not always been distinguished from the norite of the nickel intrusive, which has led to a

false picture of what the norite is. Thus, along the Southern range the norite near its outer contact always contains quartz but very little pyroxene (it has been changed to amphibole), and is usually considerably altered.

Any rock in which sudburite or recrystallization products of a similar type appropriate to its mineralogical composition occur is older than the nickel intrusive. Possibly this may serve as a criterion for distinguishing between granites younger and older than the intrusive. Older granites might be expected to show recrystallization close to the intrusive, whereas younger granites would not. Furthermore, sudburite inclusions would indicate that the granite was younger.

The formation of sudburite indicates the intense conditions prevailing during the intrusion of the intrusive. Its extent indicates the distribution of these conditions. If sudburite should be found far from the intrusive, it would indicate that it had been moved there after formation.

ACKNOWLEDGMENTS

Part of the material for this paper was obtained while the writer was acting as field assistant on the Geological Survey of Canada, under Dr. W. H. Collins, during 1928. The writer is greatly indebted to Dr. T. L. Walker for permission to use material and for other assistance. Some of the thin sections were examined at the University of Chicago.

GRAVITY SURVEY AT CROSBYTON, TEXAS

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United States Coast and Geodetic Survey

Several years ago Dr. Donald C. Barton suggested that there was an area in western Texas, in the vicinity of Crosbyton, where the torsion balance and magnetometer surveys indicated that the earth's gravitational field was strongly deformed and that in a distance of 30 miles or less the gravity anomaly might change more rapidly than in any area so far tested.[1] It was suggested that the United States Coast and Geodetic Survey occupy several stations in the area with the gravity instruments, and thus get a direct comparison between results of the pendulum instrument and those of the torsion balance and other geophysical instruments.

A survey of this area was undertaken by the Coast and Geodetic Survey in July, 1934, with the new Brown gravity apparatus, which had been perfected the previous year. Observations were made by the writer at twelve stations located in accordance with suggestions from Dr. Barton. The stations were placed from 5 to 15 miles apart on two lines passing over the structural high of the area, as determined from the torsion balance. One line was placed in a north-and-south direction and the other in an east-and-west direction.

The apparatus employed on this work is an improved form of the gravity apparatus used by the Coast and Geodetic Survey for about forty years previously to 1932. The invar pendulums and agate knife edges of the old instrument were incorporated in the design of the new one. The vacuum case was redesigned, so that the plane of support of the case is in the same horizontal plane as the knife edge, thus eliminating a large part of the flexure present in the old instrument. The new design also has an arrangement for lifting the pendulum from the knife edge and clamping it in place in the case without disturbing the vacuum seal in any way. Thus, at the beginning of a field season the pendulum is placed in the case and

[1] *Amer. Jour. Sci.*, 5th ser., Vol. XVIII (1929), pp. 154 ff.

standardized at the base station. It is then never touched, except with the starting lever, and the vacuum is not disturbed, except to pump out the small leakage that occurs, until the end of the field season and the instrument has again been restandardized at the base

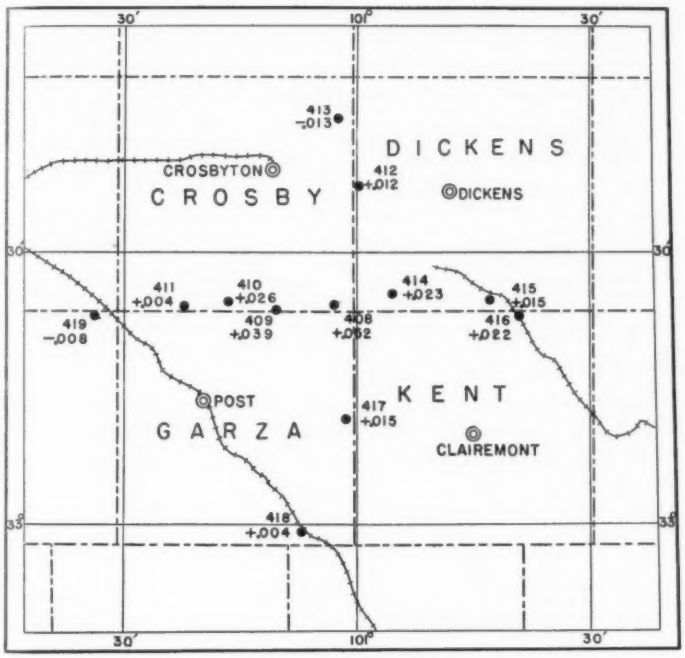

FIG. 1.—Stations in Crosbyton area, with isostatic anomalies

station. The case has been so designed that the pendulum may be demagnetized without removing it from the case or disturbing it in any way. It is believed, therefore, that one of the largest sources of error in the old observations (that of handling the pendulums between observations) has been eliminated from the new instrument. A photoelectric unit is used to pick up the motion of the pendulum and record it on a chronograph. The radio time signal is also simultaneously recorded on the same sheet, so that rates and errors of

TABLE I

PRINCIPAL FACTS AT GRAVITY STATIONS*

Station		Latitude ϕ	Longitude λ	Elevation	Observed Gravity g	Correction for—		Gravity (Bowie)‡		Anomaly (Bowie)	
No.	Name					Elevation	Topography and Compensation	Theoretical γ_0†	Computed g_c	Free-Air $g-\gamma_0''$	Isostatic $g-g_c$
		° ′ ″	° ′ ″	Meters	cm./sec.²	cm./sec.²	cm./sec.²	cm./sec.²	cm./sec.²	cm./sec.²	cm./sec.²
408	Crosbyton, No. 1, Tex.	33 24.1	101 02.8	688	979.442	−0.212	−0.001	979.603	979.390	+0.051	+0.052
409	Crosbyton, No. 2, Tex.	33 23.7	101 10.5	736	979.415	−0.227	+0.001	979.602	979.376	+0.040	+0.039
410	Crosbyton, No. 3, Tex.	33 24.6	101 16.4	747	979.399	−0.231	+0.001	979.603	979.373	+0.027	+0.026
411	Crosbyton, No. 4, Tex.	33 24.0	101 22.2	792	979.367	−0.244	+0.004	979.603	979.363	+0.008	+0.004
412	Crosbyton, No. 10, Tex.	33 37.0	100 59.8	830	979.389	−0.256	+0.013	979.620	979.377	+0.025	+0.012
413	Crosbyton, No. 9, Tex.	33 44.6	101 02.3	895	979.359	−0.276	+0.018	979.630	979.372	+0.005	−0.013
414	Crosbyton, No. 6, Tex.	33 25.5	100 55.3	705	979.413	−0.218	+0.003	979.605	979.390	+0.026	+0.023
415	Crosbyton, No. 7, Tex.	33 24.6	100 43.0	650	979.419	−0.201	+0.002	979.603	979.404	+0.017	+0.015
416	Crosbyton, No. 8, Tex.	33 22.3	100 38.6	656	979.423	−0.202	+0.003	979.600	979.401	+0.025	+0.022
417	Crosbyton, No. 11, Tex.	33 11.5	101 01.3	724	979.381	−0.223	+0.004	979.585	979.366	+0.019	+0.015
418	Crosbyton, No. 12, Tex.	32 59.3	101 07.1	707	979.357	−0.218	+0.003	979.568	979.353	+0.007	+0.004

* The new Washington (Commerce Building) base value of 980.118, on which the observed values above depend, was determined from Potsdam, Germany, in 1933.

† $\gamma_0 = 978.039[1 + 0.005294 \sin^2\phi - 0.000007 \sin^2 2\phi]$.

‡ Bowie formula of 1917.

time pieces are largely eliminated from the results of the observations.

The pendulum is swung at each station for six hours or more, and enough radio time signals are recorded during the interval so that two or more independent short-period swings may be computed for each station, in addition to the complete swing. A check on the value for each station is thus obtained without loss of time. Some of the stations in any area are reoccupied as the work progresses, to detect any possible change that may take place in the pendulum. The check between independent observations at repeat stations indicates that the accuracy of the work is about one part in one million.

The positions of the stations in the Crosbyton area, the value of the anomalies, and other principal facts regarding them are given in Table I. The locations of the stations with regard to county boundaries are shown on the sketch of the area. As was to be expected from the torsion-balance results, a rather uniform structural dome is indicated by the anomalies for the stations. No doubt the geologists interested in this area will sometime give the geological interpretation of the results of all of the geophysical surveys of this interesting region. The gravity anomalies change too rapidly over small distances to be caused in any large part by lack of isostatic equilibrium.

ICE CAVES IN NORTHERN CALIFORNIA

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University of Missouri

A recent reading of "The Origin of Ice Caves" by Harrington[1] has suggested to the writer that a brief description of certain caves in northern California may be of general interest, and that perhaps a slightly different emphasis in the explanation of the ice caves may be more satisfactory than the one given by the foregoing writer.

The caves, which are simply porous basalt-lava tubes, are in northern California in Modoc and Siskiyou counties, in and near the Lava Beds National Monument. In this region there are about three hundred lava caves of various sizes, probably one-third of which contain water or ice, or both. The water, undoubtedly of meteoric origin, has percolated downward through the porous basalt and migrated to the lower levels of the caves. If the lower ends of the lava tubes are sealed, so that the water is prevented from seeping away, a sufficient quantity may collect to be converted into ice.

Most of the ice caves in northern California are considerably larger than Shoshone Cave described by Harrington,[2] and for this reason the ice may have a different origin. Individual caves will be described to show that the angle of the sun's rays, thought to be a factor by Harrington,[3] could not play a uniformly important part in the formation of the ice. The influence of the angle of the sun's rays on caves facing north would be in contrast to that on caves facing south.

One of the best known ice caves in the region is Skull Cave. The entrance, which faces west, is about 75 feet high by 50 feet wide, and the cave is nearly half a mile long. The ice is found at the extreme end of the cave, which is perhaps 150 feet lower than the entrance.

[1] E. R. Harrington, "The Origin of Ice Caves," *Jour. Geol.*, Vol. XLII (1934), pp. 433–36.

[2] *Ibid.*, p. 434. [3] *Ibid.*, p. 435.

Frozen River Cave has a smaller entrance but is perhaps the largest ice cave in the region. The ice mass is about a hundred feet in length and of undetermined thickness. This cave faces southwest.

Crystal Cave contains more ice for its size than those described above. A feature of special interest is the abundance of hexagonal ice crystals about 6 inches long by 1 inch thick which hang from the ceiling. This cave faces east.

Bear Paw Cave, which also contains an abundance of ice, faces north.

The formation of the ice in these caves obviously is not determined by the inclination of the sun's rays at the entrance of the caves. Although the angle of inclination of the sun may be of influence in the formation of ice in caves facing south, it cannot be appealed to as an important factor in caves facing north or even those facing east or west.

Harrington points out that air circulation through fissures may be an important factor in the formation of ice in some caves. The writer is of the opinion that the presence of pronounced air currents connected with the outside through fissures would cause the comparatively free melting of the ice during the summer. Under conditions of extensive circulation of this kind, ice formed during the winter would scarcely last throughout the warm season. In some of the caves observed by the writer there is a minor amount of fissure circulation, but none is apparent in others.

The writer agrees with Harrington that the crux of the whole explanation lies in the fact that there is an active circulation in winter and little in summer. During the cold months the surface air increases in density and naturally migrates to lower levels. In doing so, it displaces warmer air that may be present.[4] This process repeats itself until the temperature at the bottom of the cave reaches or passes the freezing-point of water. During the summer months the warm surface air expands and rises, but the cold air adjacent to the ice is not displaced; thus the summer time is a season of minor circulation. The dead air spaces in the vesicular basalt act as an

[4] A. M. Miller, "Ice Caves," *Science*, N.S., Vol. XXXVII (1913), pp. 980, 981. Miller advanced a similar theory.

efficient insulator surrounding the ice on at least three sides and partially on the fourth side, protecting it from melting.

The ice in most of the caves examined by the writer does not change greatly in volume throughout the summer. Perhaps there is just enough change to bring about a balance between evaporation and melting and the formation of new ice during the winter. Ice is known to disappear from some of the caves during prolonged dry seasons, and to reappear when normal precipitation is resumed. The writer studied these caves during June and revisited them in August and noticed no appreciable change in volume. Local observers declare that the rainfall of previous years is the chief factor in determining the volume of ice present in the caves.

REVIEWS

Conodont Studies. By E. B. Branson, M. G. Mehl, and E. R. Branson. "University of Missouri Studies," Vol. VIII, Nos. 1–4. Columbia: University of Missouri, 1934. Pp. 349; pls. 28. $4.00.

During the past five years or more the authors and their students have been engaged in a comprehensive study of the conodonts. The first published results of this work include, in addition to an Introduction, six papers on Ordovician and Silurian conodont faunas, a seventh on a restudy of Hinde's British Museum type specimens, and three on basal Mississippian (?) conodont faunas.

The Introduction treats of the classification, geologic occurrence, use in stratigraphy, sedimentary associations, advantage as horizon markers, and methods employed in the study of conodonts. The authors do not regard fissile shales as likely places for conodont collecting; so "their efforts have been directed largely to shales that can be reduced easily." This is rather surprising, since the classification of conodonts now in use is largely based on forms collected from fissile shales of the Devonian and Mississippian, and in view of the fact that Ulrich and Bassler, Hibbard, Huddle, and the reviewer have collected thousands of good specimens from such rocks.

The stratigraphic studies describe conodonts from the Harding sandstone of Colorado and the following Ordovician, Silurian, and Mississippian(?) formations of Missouri: Jefferson City, Joachim, Plattin, Maquoketa-Thebes, Bainbridge, Grassy Creek, Bushberg, and Hannibal. The Hannibal fauna is described by E. R. Branson, and the others are published under the joint authorship of E. B. Branson and M. G. Mehl. The last section of No. 2 is devoted to a restudy of Hinde's specimens; and workers will find Plates 11 and 12 very useful, since they are made up of new photographs of the British Museum types. However, the reclassification of a number of Hinde's species seems without justification, and tends to confuse, rather than clarify, the situation with regard to Hinde's early work. The most striking example of this is seen in Plate 11, Figures 1 and 11, which Branson has classified as *Palmatolepis punctata* (Hinde). This species is without doubt *P. perlobata* Ulrich and Bassler, which has been found in large numbers by many workers in the Upper Devonian and

Lower Mississippian rocks. Another discrepancy is found under the description of *Hindeodella nasuta* (Hinde) in the statement that "all hindeodelloids have a large backward-directed denticle at the posterior end." The genotype *H. subtilis* Ulrich and Bassler has no such development, though a number of described species of *Hindeodella* do have such a posterior termination. The specimen referred to by Branson as *H. nasuta* was correctly redefined by Holmes as *Synprioniodina nasuta*, as shown by the attitude of the cusp and its relation to the dentition of the bar and anticusp. In referring *Lonchodus coronatus* (Hinde) to *Ligonodina ? coronata*, the assumption is made that the cusp and downward extension (anticusp) are missing. If such were the case, the missing parts would have left their impression clearly in the shale, as did the missing tips of all but one of the denticles. The form should, therefore, stand as reclassified by Holmes. *Ligonodina ? princeps* (Hinde) was redescribed by Hibbard, whose paper is not referred to in the synonomy. Neither has credit been given for the revision of *Ligonodina*.

When a complete gradational series is found in a group of fossils, it is customary to describe these forms under one species, giving the different members of the series variety names, if they are differentiated at all. Under the genus *Dichognathus* Branson and Mehl, forms from the Plattin make such a series; yet four species are described from this horizon.

It is regrettable that such a large number of fragmentary forms have been used as a basis of generic and specific descriptions, especially since "as an additional precaution species are not recognized on the presence of single specimens. With very rare exceptions, dozens, and in many cases, hundreds of specimens are available for study." It appears that, with such a wealth of material to work with, the use of fragments could be avoided. Branson and Mehl are not alone in this fault, however, since a number of authors in recent years have been publishing fragmentary forms of both conodonts and scolecodonts, which are valueless to other workers.

The fact that a number of new genera have been named for forms that might have been classified under existing genera is a debatable criticism, since the field is comparatively new and sharp lines have not been drawn with regard to generic characters. However, since the field is new, and any new fauna is bound to have a large number of new forms, as many as possible should be classified under genera already described. It is the reviewer's opinion that the genera *Bryantodus* and *Polygnathus* have been limited to a degree that makes it difficult to recognize them, and that new genera have been established on differences which heretofore have

been regarded by other workers as specific. This is especially true of *Bryantodus*, relative to which the authors have erected six new genera.

The report could have been made more usable by the insertion of the explanation of plates opposite the plates and by the inclusion of a number of corrections which should have been made in the references between plates, explanation of plates, and descriptions.

These papers make a valuable addition to the fast-growing literature on conodonts, and no doubt others will be forthcoming from this laboratory from time to time. The plates have been printed by the collotype process, which gives an excellent reproduction. In this report stereoscopic photomicrography has been used for the first time, and marks a forward step in the illustration of micropaleontological material.

CHALMER L. COOPER

Die Bodenschätze des Staates Minas Geraes, Brazilien. By B. VON FREYBERG. Stuttgart: E. Schweizerbart'sche Verlagsbuchhandlung, 1934. Pp. 453; figs. 73; pls. 12. Rm. 54.

Minas Geraes has long been famous as Brazil's great mineral-producing state. For well over a century prior to the discovery of diamonds in South Africa, Minas Geraes and Bahia supplied most of the world's diamond needs, and they are still important producers. Today the greatest known deposits of high-grade iron ore are those of central Minas. Minas is also noted for its gold, manganese, and many precious stones.

Freyberg's volume is the outcome of four trips to Brazil, during the years 1925–30, together with a thorough study of the literature of these mineral deposits. The general geology of the state is rather briefly summarized as a prelude to the economic geology which, as the title implies, is the main theme of the book. Perhaps the chief advance in general geology is greater recognition of the discordance beneath the Itacolumi quartzite, first noted by Derby in 1881. Featuring this break, Freyberg makes a separate series of the upper quartzite (Itacolumi series) distinct from the underlying iron-bearing Minas series. The latter being generally regarded as late pre-Cambrian in age, the former may be either later pre-Cambrian or early Paleozoic.

Gold has received a discussion of 81 pages, portraying in considerable detail the various occurrences and salient facts. In March, 1933, the famous Morro Velho mine reached a depth of 2,410 meters.

More remarkable are the vast beds of hematite, thus far smelted in a small way on open hearths chiefly for making knives, horseshoes, etc., but

constituting enormous future resources. These are of sedimentary origin, apparently chemical precipitates of ferric hydroxide, in places mixed with water-rounded quartz sands, but in many other places occurring as large bodies of massive specular hematite with less than 1 per cent of impurities. New factual information has been contributed by the author of this work, but our understanding of the genesis of these unusual iron ores has not been greatly advanced.

Of much interest are the 62 pages on diamonds, the extensive literature of which goes back to Jacob de Castro Sarmento (1735). The diamondiferous deposits range in age from ancient conglomerates on the broad crest of the Serra do Espinhaço (Backbone Range) to recent gravels along the present streams. Pipes of the Kimberley type not having been recognized in the Diamantina district, the ultimate source of the diamonds has occasioned much difference of opinion.

These descriptions and discussions, together with those of the many other mineral resources of Minas Geraes, make this a valuable book.

R. T. C.

Minya Gongkar, Forschungsreise ins Hochgebirge von Chinesisch Tibet. By ARNOLD HEIM. Bern: Verlag Hans Huber, 1933. Pp. 244; figs. 173; pls. 3. Fr. 9.80.

Minya Gongkar, known to Americans as Minya Konka (24,900 feet) from its first ascent by Terris Moore and R. L. Burdsall in 1932, is apparently the loftiest mountain in Western China. Though attempting no conquest of the difficult summit, the author made a journey of exploration completely around this mountain massif, by the valleys and high passes, in 1930. This exploration, together with other travels through Szechuan during 1929 and 1930, yielded much valuable geographic information interestingly presented in this attractive volume. The book is well illustrated; particularly beautiful are six plates in the natural colors.

R. T. C.

"Die Baltischen Länder (Estland, Lettland und Litauen)." By LUDWIG RÜGER, in *Handbuch der Regionalen Geologie*, Heft 28, Band IV, Abteilung 4. Heidelberg: Carl Winters, 1934. Pp. 79; figs. 14; map. Rm. 6.

This section of the *Handbuch der Regionalen Geologie* covering several small countries on the southeast shores of the Baltic treats the geology

of the region as a whole by the successive rock systems from the Proterozoic to the present. Only the Carboniferous is missing, although the Cretaceous and Tertiary systems are poorly represented. A brief discussion of the tectonic development of the eastern Baltic region is included. The mineral products are relatively unimportant.

R. T. C.

Katalog der in den Österreichischen Sammlungen Befindlichen Säugetierrests des Jungtertiärs Österreichs und der Randgebiete. By JULIUS PIA and OTTO SICKENBERG. "Denkschriften des Naturhistorischen Museums in Wien," Band 4. Geologisch-Palaeontologische Reihe 4. Leipzig und Wien: Franz Deuticke, 1934. Pp. 544. Rm. 30.

This catalogue records in systematic order all the upper Tertiary mammals of Austria and the surrounding districts. Each specimen is entered under its genus, family, and order. The following data are listed for each species: the part preserved, its age, the locality in which it was discovered, the collector, and the publication in which it has been described.

E. C. OLSON

Lehrbuch der Geologie, III Teil. *Geologische Länderkunde* (*Regionale Geologie*). By F. X. SCHAFFER. Leipzig und Wien: Franz Deuticke, 1930–34. Pp. 480; figs. 276.

Five sections of the third part of the *Lehrbuch der Geologie* by Dr. Schaffer have now appeared. Their aim is to present a regional geology of the earth. The geologic features of the continents, the ocean basins, and the geosynclinal zones are treated in great detail, considering the broad scope of the work. Regional structure is emphasized; and, wherever possible, an attempt is made to link together the major structural features of the various continents and ocean basins. The topography, stratigraphy, paleontology, and physiography of the units are treated in a more general fashion.

The text is amply supplemented by maps, stratigraphic sections, block diagrams, cross sections, photographs, and selected lists of references.

E. C. OLSON

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